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MONTHLY WEATHER REVIEW

JAMES E. CASKEY, JR., Editor

Volume 86
Number 1

JANUARY 1958

Closed March 15, 1958
Issued April 15, 1958

AN ESTIMATE OF THE MINIMUM POSSIBLE SURFACE TEMPERATURE AT THE SOUTH POLE

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[Manuscript received December 2, 1957; revised January 22, 1958]

ABSTRACT

Computations are made of the surface and tropospheric cooling which could occur during the 6-month winter night at the South Pole if only radiation-exchange processes were operating. Actual soundings taken at the beginning and end of the 1957 winter season indicate that this assumption is hardly realistic for the troposphere, but may be more applicable to the ozonosphere.

1. INTRODUCTION

The subject of extremes of temperature and other climatic elements which have been or could be experienced on the surface of the earth is one which holds an interest for most people—both scientists and laymen alike. The question of low temperatures which may occur at the South Pole during the long polar night has no doubt been the subject of much speculation. However, the establishment of a scientific station at the South Pole by the U. S. National Committee for the International Geophysical Year in January 1957, has given substance to the question. From the continuous records which are being taken, we shall at last have a quantitative measure of the annual temperature cycle, and learn the extremes to which the thermometer will dip, at least during the winter seasons of 1957 and 1958. There will, of course, always remain the question of how cold *can* it get at the South Pole.

Occurrence of an extreme low temperature near the ground at the South Pole (or at any other location) necessitates the simultaneous occurrence of an optimum combination of several meteorological elements; absence of solar radiation, clear skies, and calm air are the most essential requirements, with the ultimate fall in temperature dependent upon the duration of these conditions. In the work to be described, virtually optimum circumstances have been assumed to persist during the polar night (about 180 days), and the resulting "minimum possible" surface and tropospheric temperatures have been determined. With such background information, it is hoped that when the complete data from the IGY South Pole Station have been obtained, it may be possible to make a better quantitative analysis of the effects of the various and particular meteorological influences governing the observed wintertime temperature regimes, than would otherwise be the case.

2. THE MODEL

The computations and results which are to be described were formulated and are to be interpreted under the assumption that the snow-covered ground and the troposphere constitute a partially closed system. No external sources of energy were provided, such as would be represented by advection of warm air into the region, or a flux of sensible heat to the surface from the underlying snow. However, the upper portion of the troposphere was assumed to represent a region of heat loss in the sense that there was no downward flux of radiation into that layer which might compensate, at least in part, for the loss of radiation to outer space from the lower layers.

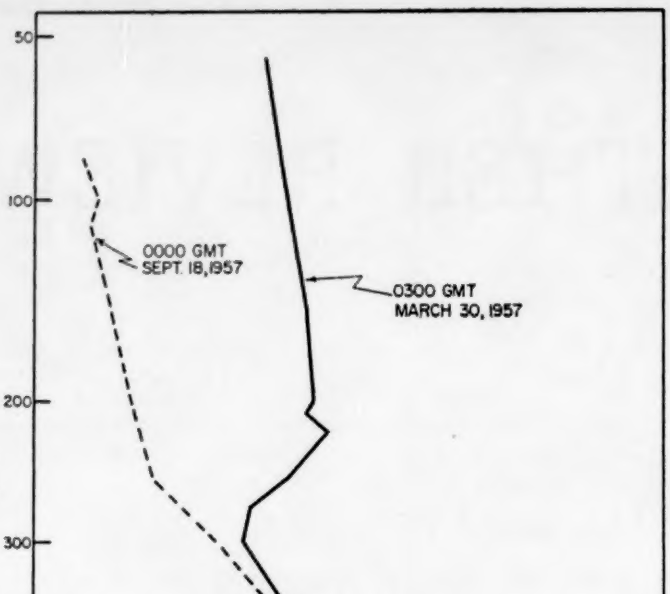

FIGURE 1.—Upper air soundings, IGY South Pole Station, at beginning and ending of the 1957 winter season.

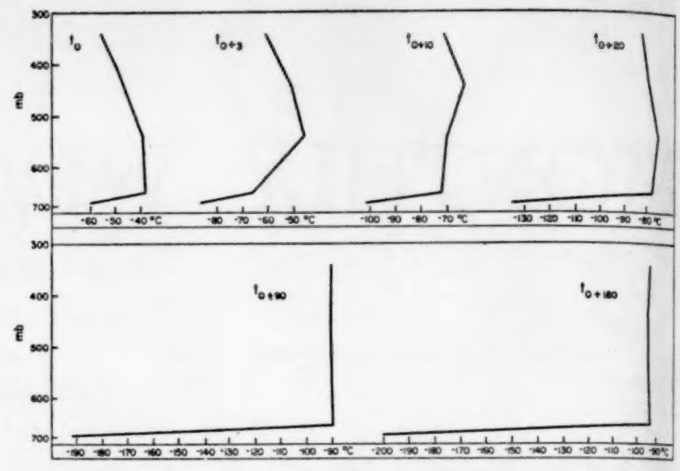

FIGURE 2.—Representative computed temperature soundings for South Pole, showing tropospheric cooling from initial time, t_0, to 180 days later, t_{0+180}.

Hence, the model is one in which, after an initial instant of time (0300 GMT, March 30, 1957; see figure 1) adjustment of the snow surface and tropospheric temperature regimes proceeds as a result of the influence of the radiation exchange processes taking place within the troposphere, between the snow surface and the inversion layer near the ground, and the radiative loss to outer space. From the initial time, and through the subsequent cooling stages, it is assumed that the troposphere is clear but saturated at all levels; that the condensed water vapor falls out of the system completely; and that the effects of the release of latent heat of condensation are negligible.

Aside from the fact that it was desired to "maximize" the tropospheric and surface cooling, the latter assumption may also be justified on the following basis. At the low temperatures which were observed in the initial sounding, the troposphere (saturated) would contain about 57.88×10^{-3} gm. of precipitable water per unit column. If during the subsequent nighttime cooling period (180 days) all of the water vapor were to be condensed out, 600 cal. $gm^{-1}\times57.88\times10^{-3}$ gm.$=34.73$ cal. would be released. This would be sufficient to warm the 670–280-mb. layer approximately

$$\frac{34.73}{(390)\,(0.24)}=+0.37°\mathrm{C}.$$

It will be seen that the computed change in the mean temperature of that layer during the period is about −45° C., or over two orders of magnitude greater.

A further simplification to reduce the number of computations involved, is to express the initial sounding as given below:

0300 GMT, March 30, 1957

Pressure (mb.)	*Temperature (°C.)*
670	−60
650	−38
540	−39
440	−47
340	−56
280	−60

As the cooling progresses, the details of the initial relative temperature distribution in the troposphere rapidly lose their identity and after the first three weeks, no vestige remains.

Since the questions to be answered cannot be expressed and solved analytically, the tropospheric and snow surface cooling was computed in a series of steps; the number of days taken in each step increased with time as the tropospheric moisture became more depleted and the cooling rates due to radiation from water vapor became smaller. Beginning from the initial time (t_0), the vertical temperature distribution in the troposphere was computed to represent the situation at 3, 10, 20, 30, 40, 50, 70, 90, 120, 150, and 180 days later. The initial and final temperature distributions as well as several intermediate ones are shown in figure 2.

3. COMPUTATIONS

The actual calculations were carried out in the following manner and approximate order:

1. Beginning at t_0, the snow surface temperature, T_s, was assumed to fall (from $-60°$ C.) very quickly to the temperature which gives quasi-equilibrium between the downcoming sky radiation (from water vapor and carbon dioxide) and the outgoing (black body) radiation from the snow surface, as described by Wexler [9]. For t_0 and subsequent synthetic soundings through t_{0+20}, the downcoming radiation, R_A, was determined by means of the Elsasser radiation chart. After t_{0+20}, however, R_A had to be estimated by some other means as the air temperatures exceeded the lower limits for which the standard Elsasser chart is constructed. A search was then made for a suitable empirical technique which might be used. It was found that within the range of R_A which could be computed from an Elsasser chart, a reasonably good linear relationship existed between log R_A and a temperature, T_m, defined:

$$T_m \equiv \frac{T_0 + T_{max}}{2} \qquad (1($$

where T_0 was the 670-mb. air temperature (°C.) and T_{max} the maximum tropospheric temperature (°C.). This relationship is shown in figure 3, and the extrapolated portion (dashed line) was utilized after t_{0+20} for determining the R_A. It will be noted that the temperature defined by equation (1) is similar to the one used by Liljequist [7] in his empirical formula for R_A as a function of T_m and vapor pressure over a range of T_m from 0° C. to $-45°$ C. Liljequist's formula was not used because it did not agree with the Elsasser chart computations in the T_m range $-50°$ C. to $-100°$ C. and also it gives negative values for R_A at vapor pressures less than 2.178×10^{-10} mb. (i. e., saturation temperatures lower than about $-140°$ C.).

2. Cooling in the inversion layer, 670–650 mb., was assumed to be made up of two components: (a) the cooling effected through the net exchange of radiation with the air layers above, described in step 3, and (b) that produced by the radiative loss to the snow surface which had taken the equilibrium temperature, T_s, described above. Hence, the equation for the cooling rate due to the latter component is

$$\frac{\partial T}{\partial t} = -\frac{(\epsilon_1+\epsilon_2)\,(\sigma \overline{T}^4 - \sigma T_s^4) \times 1440}{mc_p} \text{ °C. day}^{-1} \qquad (2)$$

where ϵ_1 = the emissivity of water vapor in the layer
ϵ_2 = the emissivity of CO_2 in the layer
σ = the Stefan-Boltzmann constant ($.825 \times 10^{-10}$ cal. cm.$^{-2}$ min.$^{-1}$ deg.$^{-4}$)
$\overline{T}$ = the mean temperature of the layer (°A.)
m = mass of air (~20.4 gm.)
c_p = specific heat of air at constant pressure (.24 cal. gm.$^{-1}$ deg.$^{-1}$)

Values used for ϵ_1, for path lengths, $w \geq 5 \times 10^{-5}$ (cm. water) were those given by Brooks [1] or extrapolated from his data. For $w < 5 \times 10^{-5}$ (encountered from t_{0+90}), an extrapolation formula

$$\epsilon_1 = 1.69 w^{1/3} \qquad (3)$$

was employed. A rigorous justification for equation (3) cannot be offered, but then neither is one necessary. It has the advantage that ϵ_1 does not become zero at very small values of w, as it does in the relation given by Elsasser [2]. In the range of w where it was applied, the emissivity is so small that it hardly makes any difference for our purposes what actual value of the parameter is used. Furthermore, the radiational cooling in the layer after t_{0+20} was quite dominated by the radiation from CO_2. For the latter a constant emissivity, ϵ_2, of 0.09 was assumed, as given by Elsasser [2] for a 150-m. layer of atmospheric air.

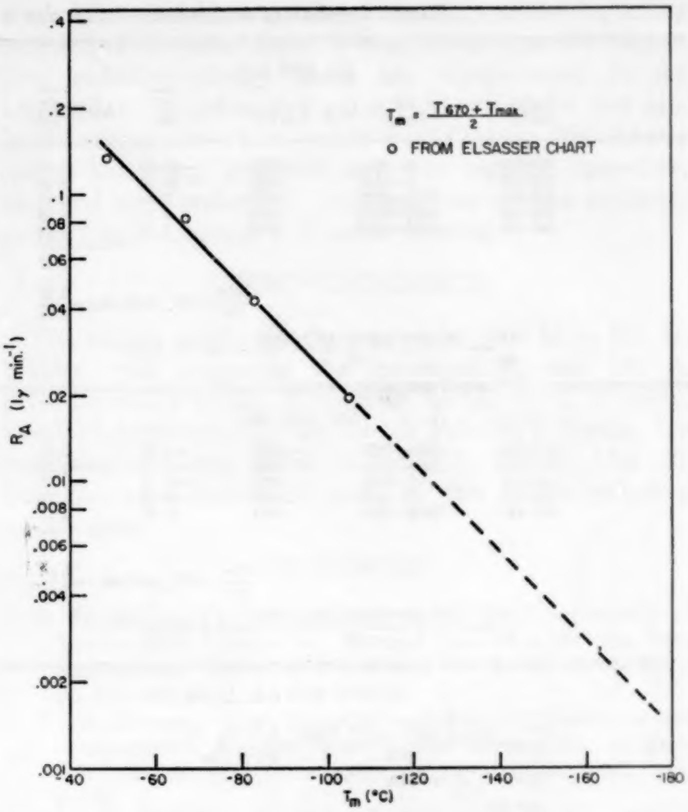

FIGURE 3.—Empirical relation between downcoming long-wave radiation and temperature, t_m.

TABLE 1.—*Computations for t_0 sounding at South Pole (0300 GMT March 30, 1957)*

p (mb.)	T (°C.)	q_s (g.kg.$^{-1}$)	$\bar{q}_s$	$-\Delta p$	$\left(\frac{\bar{p}}{1000}\right)^{1/2}$	w
670	−60	.019				
			.120	20	.812	.0014
650	−38	.220				
			.230	110	.772	.0137
540	−39	.240				
			.180	100	.700	.0088
440	−47	.120				
			.090	100	.625	.0039
340	−56	.060				
			.053	60	.557	.0012
280	−60	.046				

Note: $w = 0.7\, \bar{q}_s \frac{\Delta p}{g} \left(\frac{\bar{p}}{1000}\right)^{1/2}$

TABLE 2.—*Rate of temperature change due to water vapor at reference levels for sounding at t_0.*

(a) 650 mb.

Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\overline{\frac{\partial\epsilon}{\partial w}}$	$\Delta(\sigma T^4)\overline{\frac{\partial\epsilon}{\partial w}}$
670-650	0-.0014	−.0820	−99.90	8.19
650-540	0-.0137	.0043	20.50	0.09
540-440	.0137-.0225	.0322	3.58	0.12
440-340	.0225-.0264	.0324	2.60	0.08
340-280	.0264-.0276	.0134	2.36	0.03
				8.51

$$\frac{\partial S_E}{\partial w}(-60°, .0276 \text{ cm.}) = \frac{0.40}{8.91}$$

$$\frac{\partial T}{\partial t} = -\frac{.22\times10^{-3}\ (.806)\ (8.91)\ (1440)}{.24} = -9.5\ °\text{C. day}^{-1}$$

(b) 540 mb.

Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\overline{\frac{\partial\epsilon}{\partial w}}$	$\Delta(\sigma T^4)\overline{\frac{\partial\epsilon}{\partial w}}$
670-650	.0137-.0151	−.0820	−4.32	0.35
650-540	0-.0137	.0043	−20.50	−0.09
540-440	0-.0088	.0322	27.10	0.87
440-340	.0088-.0127	.0324	5.79	0.19
340-280	.0127-.0139	.0134	4.64	0.06
				1.38

$$\frac{\partial S_E}{\partial w}(-60°, .0139 \text{ cm.}) = \frac{0.83}{2.21}$$

$$\frac{\partial T}{\partial t} = -\frac{.24\times10^{-3}\ (.735)\ (2.21)\ (1440)}{.24} = -2.3\ °\text{C. day}^{-1}$$

(c) 440 mb.

Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\overline{\frac{\partial\epsilon}{\partial w}}$	$\Delta(\sigma T^4)\overline{\frac{\partial\epsilon}{\partial w}}$
670-650	.0225-.0239	−.0820	−2.67	0.22
650-540	.0088-.0225	.0043	−4.19	−0.02
540-440	0-.0088	.0322	−27.10	−0.87
440-340	0-.0039	.0324	49.15	1.59
340-280	.0039-.0051	.0131	12.87	0.17
				1.09

$$\frac{\partial S_E}{\partial w}(-60°, .0051 \text{ cm.}) = \frac{1.93}{3.02}$$

$$\frac{\partial T}{\partial t} = -\frac{.12\times10^{-3}\ (.663)\ (3.02)\ (1440)}{.24} = -1.4\ °\text{C. day}^{-1}$$

(d) 340 mb.

Level (mb.)	Δw (cm.)	$\Delta(\sigma T^4)$	$\overline{\frac{\partial\epsilon}{\partial w}}$	$\Delta(\sigma T^4)\overline{\frac{\partial\epsilon}{\partial w}}$
670-650	.0264-.0278	−.0820	−2.36	0.19
650-540	.0127-.0264	.0043	−3.46	−0.01
540-440	.0039-.0127	.0322	−8.26	−0.27
440-340	0-.0039	.0324	−49.15	−1.59
340-280	0-.0012	.0131	107.72	1.41
				−0.27

$$\frac{\partial S_E}{\partial w}(-60°, .0012 \text{ cm.}) = \frac{8.16}{7.89}$$

$$\frac{\partial T}{\partial t} = -\frac{.06\times10^{-3}\ (.583)\ (7.89)\ (1440)}{.24} = -1.7\ °\text{C. day}^{-1}$$

(e) 670–650 mb. layer

$$\frac{\partial T}{\partial t} = -\frac{(\epsilon_1+\epsilon_2)\ (\sigma\overline{T}^4-\sigma T_s^4)\ (1440)}{mc_p} - \frac{1}{2}\left(\frac{\partial T}{\partial t}\right)_{650}$$

$w = .0014 \therefore \epsilon_1 = .133$

$\epsilon_2 = .090$

$\overline{T} = \frac{-60-38}{2} = -49 \therefore \sigma\overline{T}^4 = .2083$

$R_A = .1334 = \sigma T_s^4$

$mc_p = (20.4)\ (.24) = 4.9$

$$\frac{\partial T}{\partial t} = -\frac{(.133+.090)\ (.2083-.1334)\ (1440)}{4.9} - 4.7 = -9.4\ °\text{C. day}^{-1}$$

(f) Sounding for t_{0+3}

Level	T (t_0)	$3\frac{\partial T}{\partial t}$	T (t_{0+3})
670	−60	--------	−87#
	(−49)	−28	(−77)
650	−38	−29	−67
540	−39	−7	−46
440	−47	−4	−51
340	−56	−5	−61
280	−60	−5*	−65

extrapolated from T_{650} and mean temperature for layer.

* assumed same as $\left(\frac{\partial T}{\partial t}\right)_{340}$.

3. Tropospheric cooling rates were determined by the tabular method described by Brooks [1] to give the rate of radiational temperature change at given levels of the atmosphere, in this instance the pressure levels listed in table 1 excepting 670 and 280 mb. Values for the 670-mb. level were found to be so near zero as to make it hardly worthwhile to compute them, and at 280 mb. there was too much uncertainty about the appropriate water distribution above that height. The equation used to determine w, the optical thickness of the water vapor atmosphere, was

$$w = -0.7(\overline{q}_s\Delta p/g)\ (\overline{p}/1000)^{1/2} \qquad (4)$$

where q_s is the saturation mixing ratio, p is pressure, and g is gravity acceleration. The "pressure-correction," $0.7\ (\overline{p}/1000)^{1/2}$, is rather a mean compromise of the factors used by various authorities; e.g., $(\overline{p}/1000)^{1/2}$ Elsasser [2], Brooks [1]; $0.7\ (\overline{p}/1000)$, Fritz [4]; $0.4\ (\overline{p}/p_0)$, Kaplan [6].

4. The cooling rates found for t_0 were applied in one step for 3 days. Then the synthetic t_{0+3} sounding was plotted, and the whole process repeated with these new data to get t_{0+10}, etc., to t_{0+180}. Selection of the number of day-units taken each time was rather arbitrary except that a temperature fall in the lowest layer of air below the T_s for the sounding was to be avoided, and some detail was desirable in order to show the transition of the sounding from its initial configuration until stabilization.

As an example, in tables 1 and 2 are given the tabulated data and computations made for t_0; the notation is virtually identical with that used by Brooks.

4. DISCUSSION

As shown in figure 2, the final condition arrived at was a snow surface temperature of about −200° C., surmounted by a strong inversion in the layer adjacent to the ground, and a sensibly isothermal layer above, in agreement with the relative temperature distribution determined for much higher temperatures by Wexler [9]. The actual occurrence of such an absolute temperature distribution at the South Pole is hardly to be anticipated. However, the magnitude of the extremity of this picture would appear to be indicative of the fact that the partially closed system model upon which the calculations were based is certainly at great variance with reality. Other factors, such as the advection of warmer air into the region and the advent of cloudiness to modify the radiation loss from the snow surface, must play important roles in the development of the snow-troposphere tem-

perature regimes during the course of the long period of darkness.

Eddy conduction of heat from the inversion will also retard the cooling at the surface, although at the expense of the internal energy of the troposphere unless replaced by advection. Liljequist's [7] investigations of this flux, Q_a, at Maudheim offer an insight into the probable order of magnitude of this term. During clear strong night-time inversions over the smooth snow fields at that location, he found the eddy flux of heat to the surface to be proportional to the wind speed at the reference level of 10 m., or,

$$Q_a=0.0058\ u_{10}\ (\text{ly min.}^{-1}) \qquad (5)$$

with u expressed in m. sec.$^{-1}$. If $u=5$ m. sec.$^{-1}$, then $Q_a=0.0290$ ly. min.$^{-1}$, compared to the computed radiative loss to the surface from water vapor and CO_2 in the inversion layer at t_0 of 0.0087 ly. min.$^{-1}$.

Some heat will also be gained by the surface during the polar night by conduction from the underlying snow. A rough estimate of the general order of magnitude of this term is also possible. From snow temperature measurements at a depth of 10 m. (where it is assumed the effects of the annual temperature cycle at the surface are negligible), Siple (communication to Wexler) places the mean annual surface temperature at about −51° C. with an annual range of the order of 67 C.°. If the classical heat-conduction model for a semi-infinite medium with a periodic variation of surface temperature (e. g., see Ingersoll et al., [5]) is applied to the snow layer (mean density and specific heat about 0.441 and 0.56 cgs, respectively) there would be a net gain at the surface of about 1800 cal., or an average of 0.007 ly. min.$^{-1}$ from the time of maximum to that of minimum surface temperatures, assumed to be 6 months apart.

In figure 1 is also shown the sounding made at 0000 GMT, September 18, 1957 near the end of the polar winter, a few hours after the new world record low temperature (at screen level) of −74.5° C. was set at the South Pole [3]. If the two soundings can be considered representative of the vertical temperature structure at the beginning and end of the polar night, it is apparent that the wintertime radiative energy losses are compensated in the troposphere by advective gains. Above about 300 mb. in the ozonosphere, however, it would appear that similar energy-balancing processes are not equally operative, as noted by Moreland [8] from analyses of data available at the Little America V Weather Central.

ACKNOWLEDGMENTS

The author would like to express his thanks to Dr. H. Wexler, who suggested the investigation, and Dr. S. Fritz for many helpful discussions, to Mr. E. C. Flowers, Chief Meteorologist of the South Pole IGY Station for supplying the data given in figure 1, and to Mrs. E. Freeman who undertook most of the rather exacting calculations.

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ON THE LOWEST TEMPERATURES ON EARTH

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[Manuscript received October 30, 1957; revised January 30, 1958]

ABSTRACT

Records on absolute minima of air temperature on the earth are analyzed geographically and in the course of time. A great deal of controversy is clarified and misleading references to incorrect values are traced to their causes. Proper documentation is provided for establishment of the true values of lowest air temperatures, as of this date, for the coldest regions on the globe—the South Polar Plateau, northeastern Siberia, Greenland, and Canada.

1. INTRODUCTION

The recent reports of record-breaking temperature minima observed in Antartica have prompted this investigation of previous minimum temperature records. Confusion has been present in the literature regarding this question and it is the aim of this paper to clarify the record. Documentation is provided to establish the true values of lowest air temperature for the coldest regions on the globe—the South Polar Plateau, northeastern Siberia, Greenland, and Canada.

2. MINIMUM TEMPERATURE ON SOUTH POLAR PLATEAU

Two successive record-breaking minimum temperatures were recorded during the 1957 winter season at the IGY Amundsen-Scott station at the South Pole. On May 11, 1957 [7] the temperature reached −100.4° F. (−73.6° C.) and near the end of the winter, on September 17, 1957 [8], a temperature of −102.1° F. (−74.5° C.) was recorded. This is, to date, the record low temperature for the world.

3. MINIMUM TEMPERATURE IN SIBERIA

The first temperature to be acknowledged as the "lowest in the world" was recorded by Neverov, a Russian merchant, on January 21, 1838, in Yakutsk (northeastern Siberia). It was observed on a Reaumur thermometer to be −48° R. (−76° F. or −60° C.). This record was found by Middendorff [18] during his travels through Siberia. It is referred to also by G. Hellmann [11]. Prior to that time, no temperature that low had been reported even by the Polar expeditions. Absolute minimum records at that time were:

Floeberg Beach	82°27′ N., 61°22′ W.	−58.8° C. (−73.8° F.)
Lady Franklin Bay	81°44′ N., 65°03′ W.	−57.1° C. (−70.8° F.)
Fort Confidence	66°44′ N., 119° W.	−57.8° C. (−72.0° F.)

In the 1890's the "Cold Pole" at Verkhoyansk (67°34′ N., 133°51′ E., elevation 107 m.) came to the fore. Unfortunately, however, the information about the lowest temperature there has been confusing from the earliest report. H. Wild, the Director of the Physical Observatory in St. Petersburg reported in Russian [32] and German [33] sources that a minimum of −68° C. (−90° F.) was recorded in Verkhoyansk on January 15, 1885. He added the confusing remark: "Should we reduce the reading of the alcohol thermometer to the air thermometer, we would obtain, for the above mentioned minimum, a temperature of −76° C." (−105° F.). There was no explanation for the reduction, or why so great a difference would be obtained. The Russian Meteorological Yearbook (Annales) for 1885 shows the minimum for January of that year as −67.1° C. (−88.8° F.). In view of this and the lack of explanation for Wild's reduction, we must discard Wild's figure of −76° C. (−105° F.) and accept the figure given in the Yearbook as the correct minimum for that year.

The next figure to appear in the literature as the "lowest in the world" was again at Verkhoyansk, in February 1892, when the absolute minimum for the month was observed on the 5th and again on the 7th as −69.8° C. (−93.6° F.). This value has been frequently quoted as the lowest temperature on record. It was, however, later corrected, as we shall see below.

In 1910 A. I. Voeĭkov [28] gave the minimum temperature at Verkhoyansk as −72° C. (−97.6° F.) for February 1892. Since this represented the "lowest temperature on earth" and was at variance with previous records, Prof. C. F. Marvin, Chief of the U. S. Weather Bureau at that time, requested an explanation of the data from Prince Boris Golitsyn at the Nicholas Central Physical Observatory, Petrograd. The answer from Prince Golitsyn [9] explained that neither the minimum of −69.8° C. (−93.6° F.) nor the figure of −72° C. (−97.6° F.) was correct. It suggested that Voeĭkov's error was most probably due to the fact that a correction of −2.0° C. was added once more to the already corrected reading of the alcohol thermometer published in the Yearbook. The letter concluded, "The lowest temperature of air in

Verkhoyansk ought therefore to be considered as −68° C. (−90.4° F.) it is also the lowest temperature which has ever been observed on the stations of our meteorological net."

Thus, the round figure of −68° C. was established as the "lowest on earth", and has been generally used in Russian textbooks since. The erroneous figure of −69.8° C. has unfortunately continued to appear also, and this prompted E. Rubinshtein [22] to publish another explanatory article. She repeated the explanation given by Golitsyn and concluded: "The direct reading of the spirit thermometer on February 5 and 7, 1892, in Verkhoyansk was −67.8°. Adding the instrumental correction of +0.2°, we obtain −67.6° which, in our opinion, must be taken as the most probable thermometer reading of these days."

An expedition, sponsored by the Akademy of Sciences and headed by the geologist Sergei Obruchev, was sent to the Oimekon region, also in northeastern Siberia, in 1926, and discovered there a second cold pole. In his first communication [19] he remarked:

> Temperatures of −50° to −60° C., which set in by the end of November, greatly surprised us, for thus far it was supposed that the "Pole of Cold" (with temperatures to −69° C.)* was at Verkhoyansk; our observations permit the extension of the cold area as far as Oimekon, and possibly the transference of the pole itself to the latter place; but the solution of this question will only be possible after systematic observations during the whole winter.

Karl H. Pollog [21] mentions that Obruchev measured a temperature of −60° C. in November near Oimekon. In a second article, Sergei Obruchev [20] states:

> First of all, in order to avoid any misunderstandings, I must say that, up to this time, I have not published any exact data on temperature and those that appeared in my article were only some approximate figures. Our expedition was not supposed to take meteorological observations, and since it was planned to return to Yakutsk before winter set on, we had taken along only the mercury thermometers. Therefore, we could not measure any temperatures below −39.4° C. (mercury freezing point) . . . we had to make use of some objective indications—namely of the rustling sound produced by the freezing breath, which is much like a sound of pouring grain. This phenomenon is called by the Yakuts the "whisper of stars." According to observations of Cherskii it appears at temperatures of −48.5° C. and lower. During the period of very low temperatures, from the 10th to the 21st of November, I have heard this sound beginning from 6 p. m. on. This fact provided a reason why I was telling about the temperature as being from −50° to −60° (and not exactly −60°, as it has been mentioned by C. Pollog).

Unfortunately it appears that the −60° C. reference by Pollog was converted to −76° F. and later appeared in the European literature as simply −76° without reference to scale. An anonymous report [17] presumably communicating the results of studies by Prof. Sandström, speaks of the "frightening temperature of 76° below zero" at Verkhoyansk and says of Oimekon that "Professor Cherskii recorded the alarming temperature of 78° below zero at this desolated site." One can only assume that these temperatures, mentioned without reference to scale,

*Obruchev is here using the wrong figure for Verkhoyansk.

Table 1.—*Absolute minima of temperature (°C.)*

	Verkhoyansk	Oimekon		Verkhoyansk	Oimekon
January	−67.2	−65.6	July	−2.3	−4.4
February	−67.6	−67.7	August	−7.9	−7.1
March	−60.3	−56.7	September	−16.7	−17.8
April	−54.5	−46.3	October	−44.6	−35.7
May	−28.1	−25.4	November	−56.6	−62.2
June	−7.3	−5.8	December	−64.5	−64.4
			Year	−67.6	−67.7

are meant to be Fahrenheit readings. Published as they were in European publications, however, they were considered by other authors to be in °C. They were quoted and requoted in various publications and gradually became established in the literature [6, 12, 15]. Since the observation at Oimekon was attributed to Cherskii, one can review his reports to verify the data. Such an examination, however, fails to reveal a value which could be interpreted as −78° in any scale. Cherskii wintered in northeastern Siberia in 1891–92 and in his report [4] made only sparse remarks about the temperature. The lowest value which he mentions is −58° C.

If we assume, as we must, that the reference to −78° is erroneous, we must then determine what the true "lowest" temperature observed at Oimekon was. We know, that following Obruchev's suggestion, a meteorological station was set up in Oimekon in 1929. Since then some data obtained from observations taken by this station have been published in Russia, by prominent authors, presenting a reliable and also an official source of information. The greatest scope of data for Oimekon is found in the "Climatological Handbook for the Soviet Sector of the Arctic," Leningrad, 1940, edited by E. I. Tikhomirov [26]. In table 5 of this source are given the absolute minima of temperature for Verkhoyansk and Oimekon, which are reproduced here in table 1. These data were obtained from the actual series of observations for the period 1891–1920 in Verkhoyansk, and for 1931–1935 in Oimekon.

As we can see, even though Oimekon has a much shorter series of observations, it still came out with a little lower minimum than Verkhoyansk, and consequently, the title of the "Cold Pole" has been won by Oimekon.

In the same source we find also the mean monthly temperatures, year by year, and we reproduce them here (table 2) for the years 1931–1935, when both stations were working, for the colder months. As can be seen from table 2, the differences in the mean monthly temperatures between Oimekon and Verkhoyansk can be very great indeed; in single months Oimekon had a mean temperature lower by 10° to 12° C. than had Verkhoyansk.

If we compare the mean monthly temperature for the whole period of 5 years (1931–1935) for which we have the parallel observations of both stations, we obtain the differences (C.°) in mean monthly temperatures between Oimekon and Verkhoyansk given in table 3. Only November was consistently warmer in Oimekon, but all the other months were colder.

TABLE 2.—*Mean monthly and annual temperature (C.°)*

Years	January		February		March		December		Year	
	Oimekon	Verkhoyansk	Oimekon	Verkhoyansk	Oimekon	Verkhoyansk	Oimekon	Verkhoyansk	Oimekon	Verkhoyansk
1931	-------	−50.1	−49.5	−46.9	−27.5	−29.6	−50.3	−39.5	-------	−14.2
1932	−50.9	−48.7	−45.6	−38.9	−32.5	−28.5	−49.6	−49.0	−16.8	−15.2
1933	−53.2	−40.6	−47.4	−47.8	−32.8	−33.5	−41.5	−46.8	−15.5	−14.6
1934	−47.1	−45.7	−44.3	−41.6	−33.6	−31.9	−47.6	−47.2	-------	−14.9
1935	−50.8	−49.4	−49.1	−40.8	−34.9	−34.1	−42.8	−40.2	−15.4	−14.5

TABLE 3.—*Difference between mean monthly temperatures (C.°) at Oimekon and Verkhoyansk, for period 1931–35. Minus sign equals colder in Oimekon.*

January	−4.4	July	−1.9
February	−4.1	August	−1.1
March	−0.8	September	−1.1
April	−1.3	October	−1.6
May	−0.9	November	+1.5
June	−1.5	December	−1.9
		Year	−1.7

The absolute minima of −67.6° C. (−89.7° F.) for Verkhoyansk, and −67.7° C. (−89.9° F.) for Oimekon are also given in a later, and very important, work by Vize [27] which confirms their correctness.

The first one who published Oimekon's lowest temperature of −67.7° C. was Salishchev [23]. In his article he gave a graph of the monthly mean, maximum, and minimum temperatures in Oimekon for the years 1930–1934, and stated: ". . . an absolute minimum for the four years of −67.7° C. (−89.9° F.) was recorded on February 6, 1933." It is interesting to note the close coincidence of dates when the lowest temperatures were recorded at both places. In Verkhoyansk it was on the 5th and 7th day of February (1892), in Oimekon on the 6th of February (1933). Apparently the first 10 days of this month might be regarded as the most dangerous period in respect to cold in this area.

Most recently the correct lowest temperatures for Verkhoyansk and Oimekon were published in the United States by Stepanova [25].

Of course, scientists may wonder whether Russians have recorded some lower temperatures in recent years. A profound search through all Russian periodicals and most recent textbooks in meteorology has shown that the illegitimate figures of −76° and −78° have never crossed the Russian border and have never appeared in Russian sources. In the most recent textbooks on climatology, which had been approved and recommended for use in universities, the same old, round figure of −68° C. (established by Golitsyn) still represents the lowest temperature on earth. Kostin and Pokrovskaiâ [14] gave the same value as the lowest temperature ever observed in the "Verkhoyansk-Oimekon Cold Pole region." Another recent textbook by Alisov, Berlin, and Mikhel' [1] is still more conservative; on p. 65 it says:

> In the coldest years 1885 and 1892, the lowest temperature on the earth's surface was recorded in Verkhoyansk −68°. In Oimekon, which is located on the plateau about 700 m. above sea level, the absolute minimum of air temperature might be still lower.

So, potentially, the Oimekon region has indeed all the advantages for originating the lowest temperatures. The orographic map in figure 1 shows the inner Oimekon Plateau as being a more elevated and at the same time more protected area than that of the Verkhoyansk region. Greater elevation stipulates a stronger cooling effect of the outgoing radiation since the re-radiation from the atmosphere usually decreases with height under the clear sky. But considering the fact that at low temperatures the outgoing radiation is generally small, the second factor weighs much more in the explanation of the strong cooling. The role of the mountain barrier encompassing the Oimekon Plateau and creating the extremely calm atmosphere over it seems to be of paramount importance in the origin of the lowest temperatures in this place. This conclusion is in accord with the evidence from the South Pole which indicates that the occurrence of the extremely low temperatures is primarily due to the establishment of such conditions where minimum mixing in the surface layer that contains the inversion is attained (Wexler [30]).

4. MINIMUM TEMPERATURE IN GREENLAND

In comparing the lowest temperatures of the Greenland Ice Cap, it was found that the absolute minimum at "Eismitte" (71.2° N., 39.9° W., 2993 m.) as recorded by A. Wegener's [29] expedition on March 20, 1931, was only −64.8° C. (−84.6° F.). F. Loewe [16] reported that ". . . the lowest temperature recorded on the Greenland ice sheet was −66° C. (−87° F.) on 21 February 1950 at the French central station in lat. 70°54′ N., long. 40°42′ W., 2993 m. (9820 ft.)." However, on page 96 of the French publication of observations of the French station in Greenland [5], a temperature of −64.8° C. (−84.6° F.) is found for February 22, 1950 (which is 9 p. m. of February 21, in Greenland). The coordinates are slightly different and there are no remarks concerning a correction of coordinates. The height is the same. It must therefore be assumed that these are the best data (that are published in the station record), and that the lowest temperature observed by the Paul-Émile Victor expedition is exactly the same (−64.8° C. (−84° F.)) as that recorded during A. Wegener's expedition.

5. MINIMUM TEMPERATURE IN CANADA

For Canada, the official lowest temperature is −81° F (−62.8° C.). It was observed on February 3, 1947, in

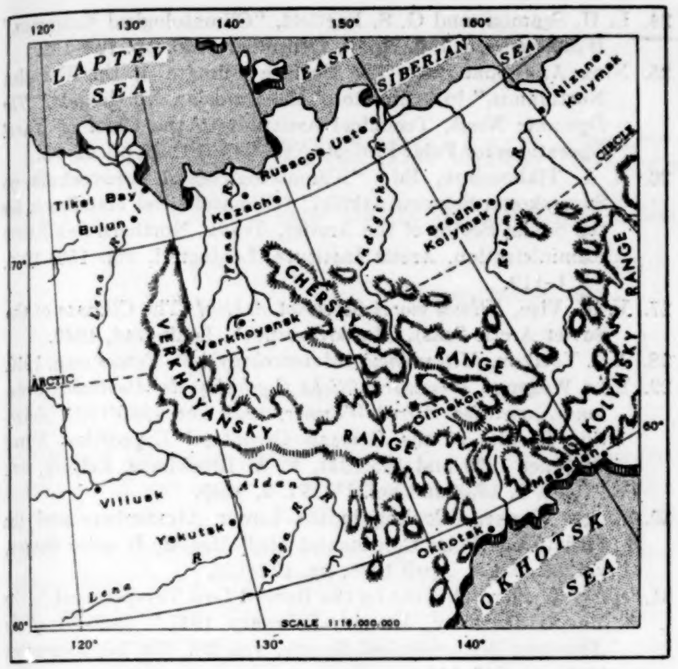

FIGURE 1.—Map of northeastern Siberia, showing the locations of Verkhoyansk and Oimekon.

the Yukon Territory at Snag, 62°23′ N., 140°23′ W., elevation 2,120 feet (646 m.).

This absolute minimum was published in the official source of observation data *Monthly Record* [2], and was used in recent studies by Kendrew and Kerr [13]. It was also applied by Hare [10], who claimed, however, that the authenticated record was −81.5° F. (showing higher accuracy), but gave no evidence and no reference to the original source. Seamon and Bartlett [24] have investigated the meteorological extremes and have also found that the lowest official temperature ever recorded in Canada was −81° F. at Snag on February 3, 1947.

However, it should be mentioned here that a different minimum of air temperature was published in the *Monthly Weather Map* [3]. In tables on the reverse side of the map for February 1947, an absolute minimum of −83° F. is indicated for Snag, and the text of this issue states: "At Snag, Yukon Territory, a minimum temperature of −83° F. was reported which set a new low record for all Canada."

Thus in Canada, we have two official sources with conflicting data. An explanation of this fact is found in an article by Wexler [31]. A footnote in the article states that

> . . . the lowest graduation on the Snag thermometer was −80° F. and by extrapolation −83° was observed to be the minimum. The official Canadian records indicate the extrapolation was actually to −84° F., but the Canadian authorities recognized −81° F. as the official minimum after having calibrated the thermometer and found an instrumental error of +3.0° at that temperature.

The previous Canadian record low temperature was −78.5° F. (−61.4° C.) observed at Ft. Good Hope [31] on December 30, 1910.

6. CONCLUSION

Thus, it has been established that there are four very cold regions on our planet: Yukon Territory, with the lowest temperature −81° F. observed at a height of 646 m.; Greenland, with the lowest temperature −84.6° F. observed at height 2993 m.; Oimekon, with lowest temperature −89.9° F. at height of about 800 m.; and the South Polar Plateau with the record breaking temperature of −102.1° F. at height of about 2800 m.

ACKNOWLEDGMENTS

The author is indebted to Dr. Harry Wexler for suggesting this investigation; to Dr. Helmut Landsberg for encouragement and advice; to Mr. William Haggard and Miss Gertrude Fricke for editorial assistance; and to Mr. Paulus Putnins who contributed the information on temperatures in Greenland.

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THE WEATHER AND CIRCULATION OF JANUARY 1958[1]

Low Index with Record Cold in Southeastern United States

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1. HIGHLIGHTS

January 1958 was characterized by a major southward migration of the upper-level westerlies into the subtropics (an index cycle) in the Western Hemisphere. Near mid-month an almost record-breaking blocking anticyclone consolidated in Davis Strait, causing a succession of "northeasters" to stagnate off New England. These conditions persistently deployed cold polar air into the southeastern United States.

Florida was probably the hardest hit section, with at least seven active cold fronts replenishing the cold air and bringing exceptionally heavy rains, and even some snow, to northern and central sections. This added up to one of the most disastrous Januarys for a large portion of Florida's economy. The citrus output was reduced by millions of boxes; damage to citrus trees may require 2 or 3 years to repair; dead pasture grasses, in addition to the cold and wet weather, killed hundreds of head of cattle; and tourist trade income was sharply reduced.

The same anomalous circulation regime brought much above normal temperatures to parts of New England as persistent easterly flow brought relatively mild maritime air inland. This was associated with the heaviest precipitation of record in some places; e. g., Boston reported 9.54 inches for the month, more than double the normal amount.

Persistent ridge conditions aloft in the western portions of North America produced above normal temperatures from the Central Plains and Upper Mississippi Valley westward to the Pacific coast, with many places enjoying the second or third warmest January of record. For example, Glasgow, Mont., reported this was the only January that temperatures did not fall below zero, and Meacham, Oreg., had the warmest January on record.

Fogginess was much above normal in the Great Central Valley of California due to the persistence of a strong Great Basin anticyclone. Sacramento, Calif., reported 19 days of heavy fog or 3 times the normal, and Red Bluff, Calif., reported 8 days of fog, equal to the normal for a whole winter.

At low latitudes there were many other interesting highlights attributable to the fact that the Western Hemisphere jet stream was depressed far south of its normal location. Swan Island in the Caribbean reported 3 "northwesters" with fresh to strong winds 3 to 5 days after passage. In the Pacific, practically all Hawaiian stations reported below normal precipitation, with Hilo recording a total of only 2.91 inches, or 11 inches below normal.

[1] See Charts I-XVII following p. 40 for analyzed climatological data for the month.

2. GENERAL CIRCULATION

The 700-mb. mean circulation for the month (fig. 1) was quite similar to that of January 1956 [1] except for one major difference, the absence of a block in the Bering Sea this January. This difference may have been, at least in part, responsible for the difference in temperature anomalies in the northern Plains between the 2 years. This January temperatures averaged well above normal, whereas in 1956 temperatures were below normal.

Three noteworthy abnormalities developed in the mean circulation over the Western Hemisphere this January:

1. A trough deepened in the east-central Pacific with 700-mb. mean heights over 500 feet below normal about 700 miles south of Kodiak, Alaska (fig. 1), and sea level pressures 20 mb. below normal (Chart XI inset). The abnormally deep trough in this position maintained a persistent ridge aloft, and an associated Great Basin High at the surface, in the western United States. This east-central Pacific trough represented a marked eastward displacement of the trough normally just off the eastern Asiatic littoral. In addition a stronger than normal subtropical anticyclone in the western Pacific, directly south of a slightly deeper than normal Low in Kamchatka, resulted in a stronger than normal confluence of westerlies near southern Japan. Vorticity maxima which developed along the strong Japanese baroclinic zone continually propagated eastward to maintain the east-central Pacific trough and center of action. (Note the 986-mb. Low just south of Kodiak on the monthly mean sea level map, Chart XI.)

2. The second noteworthy abnormality was the block in the Davis Strait where monthly mean 700-mb. heights averaged 520 feet above normal (fig. 1). This block predominated during the second half of the month, when it developed to near record proportions.

3. The third abnormality was the center of below normal heights in the southeastern United States, which persisted through the entire month and was accompanied by northerly surface flow on the average over the eastern, and particularly the southeastern, United States. The stronger

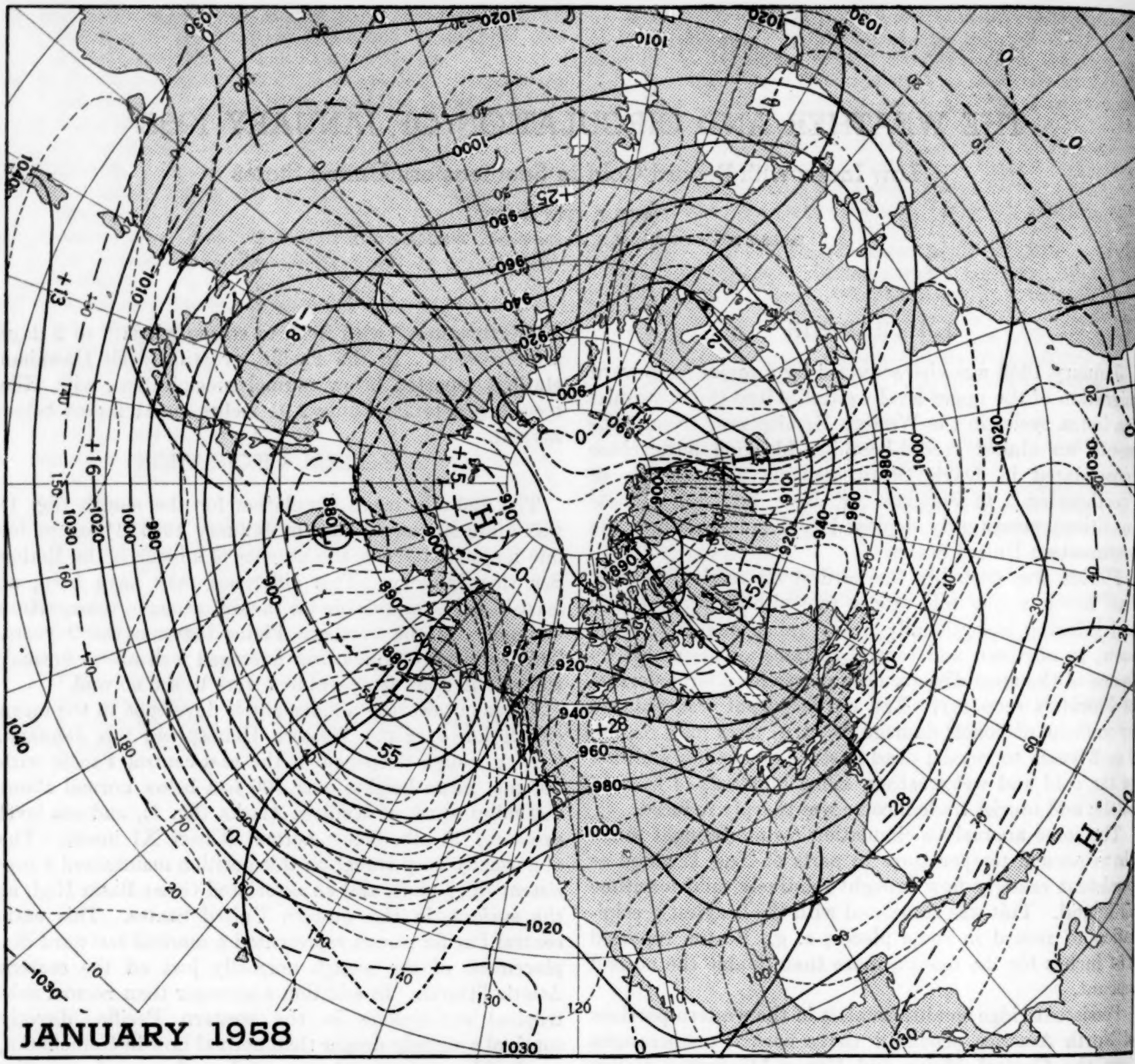

FIGURE 1.—Monthly mean 700-mb. contours for January 1958 labeled in tens of feet, and height departures from normal (short dashed lines at 50-ft. intervals, with centers labeled in tens of feet and zero isopleth heavier). Trough lines (heavier solid lines) connect minimum latitudes of contours.

than normal westerlies between 25° N. and 40° N., particularly in the eastern Pacific (fig. 2), may have forced the strongly sheared and positively tilted trough, which normally extends from the St. Lawrence Valley southwestward across Baja California in January, eastward to the middle Atlantic coast and southwestward across the northwestern Gulf of Mexico. This contributed to the strong negative height anomaly over the southeastern United States. This anomaly, coupled with the blocking in the Davis Strait, was of great importance in determining the monthly weather regime in the eastern half of the United States.

An abrupt change in the circulation took place near the middle of the month when the zonal westerlies in the Western Hemisphere (fig. 3) began to decline at an increasing rate about January 11. At this time continuation and intensification of a strong index cycle [2] was indicated by rapid anticyclogenesis in the Davis Strait area. This is more clearly illustrated by figure 4, which shows the two 15-day averages which constituted the monthly mean 700-mb. circulation. The negative height departures from normal for both halves of the month were quite similar in the eastern Pacific and the south-

eastern United States. However, the biggest change occurred in the Greenland-Davis Strait area where below normal heights in the first half of the month were replaced by a tremendous block averaging over 1,000 feet above normal in the second half. This block resulted from a consolidation of the two positive height anomaly centers observed on the first 15-day mean map (fig. 4A) as the Atlantic center retrograded to the Davis Strait and combined with the northeastward-moving center from the Alberta-Montana area. This consolidation was closely associated with an intensification and retrogression of below normal conditions in the southern and eastern United States.

On a 5-day mean basis, the Davis Strait block reached its maximum intensity at 700 mb. of about 1,500 feet above normal in the January 18–22 period. On a hemispheric basis, this was probably second in intensity only to the record anomaly of +1,600 feet in the same area during the 5-day period February 19–23, 1947.

3. CHANGE IN CIRCULATION FROM DECEMBER, AND ONSET OF INDEX CYCLE

The circulation during the previous month, December 1957, was one of high index and abnormal warmth in the United States [3], except for one major cold wave which produced the first severe winter freeze in Florida early in the month. In addition, December was characterized by the fact that the mean westerlies in the Western Hemisphere were stronger and farther north than normal, while a strong positive height anomaly (or block) persisted just off Newfoundland. Both of these circumstances have been recognized as conditions antecedent to an index cycle, according to Namias [2], provided that a reservoir of Arctic air in the polar regions is sufficiently great. Thickness anomaly charts during the last week of December (not shown) indicated that this reservoir of cold air was quite extensive and intense over Canada. Although late December is somewhat early for the beginning of a major index cycle, and although such a development would seem to conflict with the stronger persistence found between December and January than between any other pair of months [4], nevertheless the recognized prerequisite conditions seemed to be fulfilled at this time. Furthermore, a recent precedent for a strong index cycle this early in the winter was set in January 1956 [1], although that case represented persistence of low index from its preceding December.

Figure 3 shows the variation with time of overlapping 5-day mean 700-mb. wind speeds for both the temperate westerlies and subtropical westerlies during December 1957 and January 1958. It can be seen that about December 20 the subtropical westerlies started to increase, while the temperate westerlies went into decline. This was the start of a major prolonged index cycle which continued through the entire month of January with no recovery evident at month's end. The extensive cold outbreak which swept southward across the United States in the first days of January was one of the early manifestations of this developing index cycle.

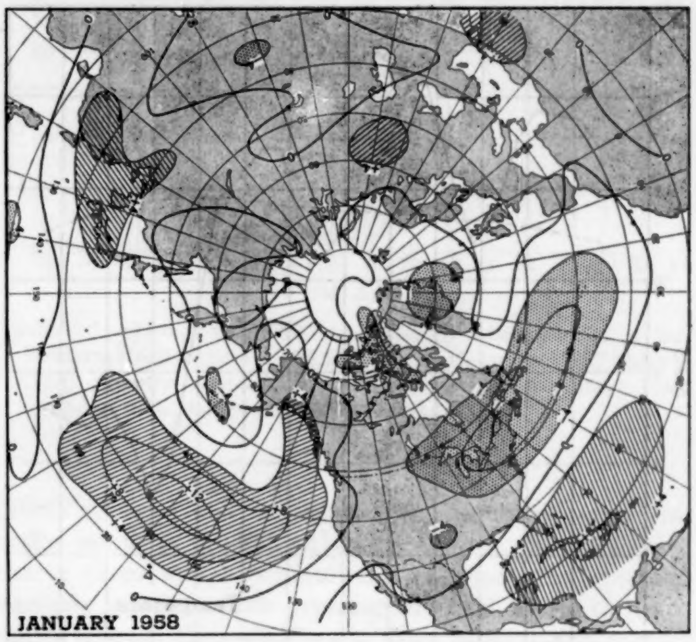

FIGURE 2.—Departure from normal of monthly mean 700-mb. wind speed (meters per second) for January 1958. Hatched areas indicate more than 4 m. p. s. above normal; dotted areas, more than 4 m. p. s. below normal.

Figure 5 shows how the latitudinal profile of 15-day mean westerly wind speeds in the Western Hemisphere changed with time. The high speeds in temperate latitudes during the last 15 days of December declined sharply by the last half of January. The sharp peak of near 14 meters per second at about 32° N. in the latter half of January presents a striking contrast to the peak speed of similar magnitude at a much higher latitude during the latter half of December; i. e., the hemispheric jet axis migrated about 15° southward during the period.

Another graphical illustration of the startling behavior of the circulation associated with this index cycle is given in figure 6. Starting with the last above normal peak wind speed of about 16 m. p. s. on the 20th of December 1957 near 47° N., the 700-mb. hemispheric jet axis migrated southward until the early days of January when the axis reached 40° N. For the remainder of the first half of the month an attempt at a recovery occurred during which time warmer air progressed across the United States and confined the below normal temperatures to the Southeast. Referring back to figure 3, it can be seen that although both temperate and subtropical westerlies averaged near normal during this period, the subtropical westerlies were continuing a steady trend upward. The fact that the westerlies averaged near normal during the first half of January could be misleading as to the state of the circulation, since the steadily rising subtropical westerlies indicated continued intensification of the index cycle, while the temperate westerlies were undergoing an abortive attempt at recovery.

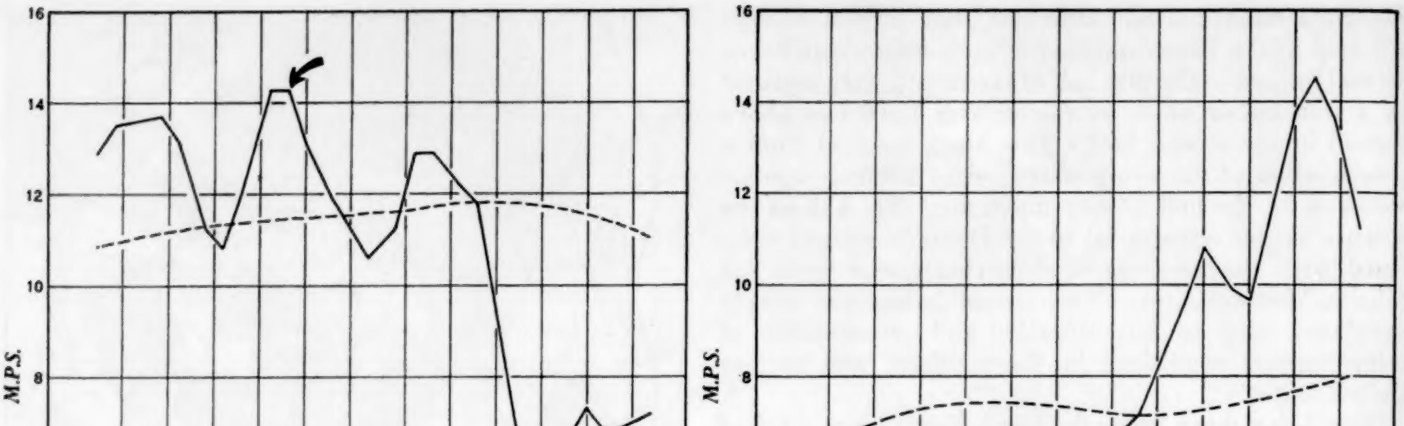

FIGURE 3.—Time variation over the Western Hemisphere of 5-day averages of 700-mb. westerlies (m. p. s., solid lines) and the normal values (dashed) for (A) 35°–55° N. (temperate) and (B) 20°–35° N. (subtropical.) Heavy arrows indicate start of index cycle.

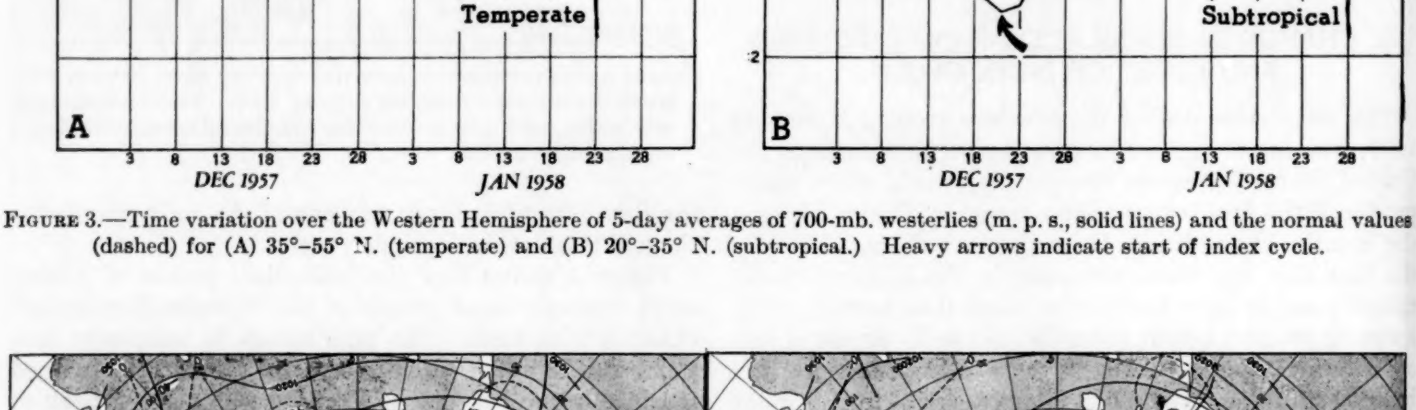

FIGURE 4.—15-day mean 700-mb. contours for the two halves of January 1958 with height departures from normal superimposed as dashed lines (every 200 ft.). Heavy dotted tracks represent smoothed paths of 5-day mean positive anomaly centers on (A) which consolidated to form Davis Strait block on (B).

Thus, during the middle weeks of January, multiple axes of strong westerlies (fig. 6) diminished at middle and higher latitudes as the jet axis in the lower latitudes continued to increase in strength, reaching a peak speed, and perhaps a record value, of 17.1 m. p. s. for two successive overlapping 5-day mean maps at 33° N. near the end of the month.

To complete the picture of this month's circulation, figure 7 shows the mean 200-mb. contours with mean isotachs and jet axes superimposed. At this level the jet stream was located in the northern Gulf of Mexico, in contrast to the previous month when it was located through Tennessee. The jet axis was split over the Atlantic, with the northern branch reflecting the influence of the Atlantic block primarily in the first part of the month, and the southern branch responding to the southward migration of the mid-latitude westerlies in the second half of the month. Two well-defined speed maxima existed in the Northern Hemisphere, one of about 71 m. p. s. off southern Japan, and another of about 53 m. p. s. in the Gulf of Mexico.

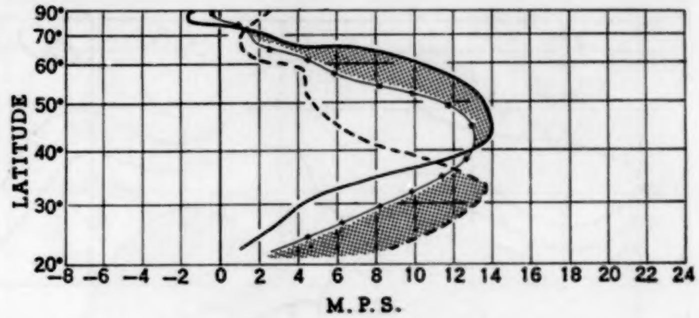

FIGURE 5.—15-day mean 700-mb. westerly wind profiles in the Western Hemisphere for the last half of December 1957 (heavy solid), January normal (thin solid), and last half of January 1958 (dashed), with above normal speeds shaded.

4. CYCLONE AND ANTICYCLONE TRACKS IN THE UNITED STATES

Two types of storms dominated the monthly weather picture. Undoubtedly the most important were the "Gulf" types, some of which moved northeastward along an inland path, but the majority became "northeasters" following a more normal track northeastward off the Atlantic coast, as shown in Chart X. The monthly mean sea level map (Chart XI) shows a trough off the east coast, reflecting the path of many deepening storms, in agreement with the position of the mean 700-mb. trough along the east coast in figure 1. The Low on the mean sea level chart southeast of Nova Scotia reflects the stagnation of the deep daily Lows in this area which had such a strong influence on the weather in the northeastern United States.

A weak trough on the monthly mean sea level chart from Alberta through southeastern Montana reflects a secondary locus of "Alberta" type disturbances which were generally weak and tended to fill as they moved eastward and southeastward. Due to the influence of blocking conditions in eastern Canada, these storms failed to recurve northeastward and deepen in the usual fashion.

The strong High on the mean sea level chart in the western United States, averaging about 1024 mb. for the month, reflects the relative absence of surface cyclonic systems of any consequence crossing the western Plateau, and also the frequency of daily anticyclones in this area (Chart IX). No single preferred anticyclone path appeared in the eastern United States, with Canadian Highs tending to "glance" eastward north of the Great Lakes toward Labrador in the latter part of the month, and another favored locus lying across the northern Gulf of Mexico (Chart IX).

Tracks of major Pacific cyclonic systems for the most part turned northward before striking the west coast of North America and contributed to the strong center of action south of Kodiak, Alaska. On the daily 500-mb. charts there was a predominant tendency for cyclonic vorticity maxima, after crossing the west coast, to drop southeastward through Nevada, New Mexico, and northern Texas, where surface cyclonic activity was intensified or regenerated along the polar front in the northwestern Gulf of Mexico. This persistent tendency is, of course, reflected in the location of the monthly mean trough aloft in the northwestern Gulf. The jet maximum at 200 mb. in the Gulf of Mexico (fig. 7) reflects the preference of this region as a locus of vorticity centers and sea level cyclones.

5. INDIVIDUAL STORMS

During the first week in January a strong cold anticyclone crossed the United States producing below normal temperatures east of the Plains. Progress of this cold High was retarded on the 3d by the worst winter storm in Miami's weather history. This storm formed near Cuba and hit southeastern Florida, with winds gusting to 70 m. p. h. and 3 to 5 inches of rain. This activity was related to the very early stages of the index cycle.

On the 5th and 6th another low-latitude storm in northern Mexico caused flooding rains of 6 inches or more in the lower Rio Grande Valley and 7 inches of snow in extreme western Texas and eastern New Mexico. This cyclone moved across the northern Gulf and up the Atlantic seaboard with heavy rains enroute. At Nantucket early on the 8th this storm deepened to a record low barometer of about 960 mb., with winds of hurricane force near the center and heavy snows along a 50-mile belt from Virginia northward. Storrs, Conn., for example, collected 17 inches of snow, a new record.

Persistent northerly winds along the east coast in the storm's wake drove down weekly average temperatures 10° to 13° below normal over Florida and the Southeast during the second week, and damaged citrus as temperature minima ranged from the low 20's to the 30's on the 9th.

Another storm formed in southern Texas on the 12th and moved slowly to southern New England on the 14th

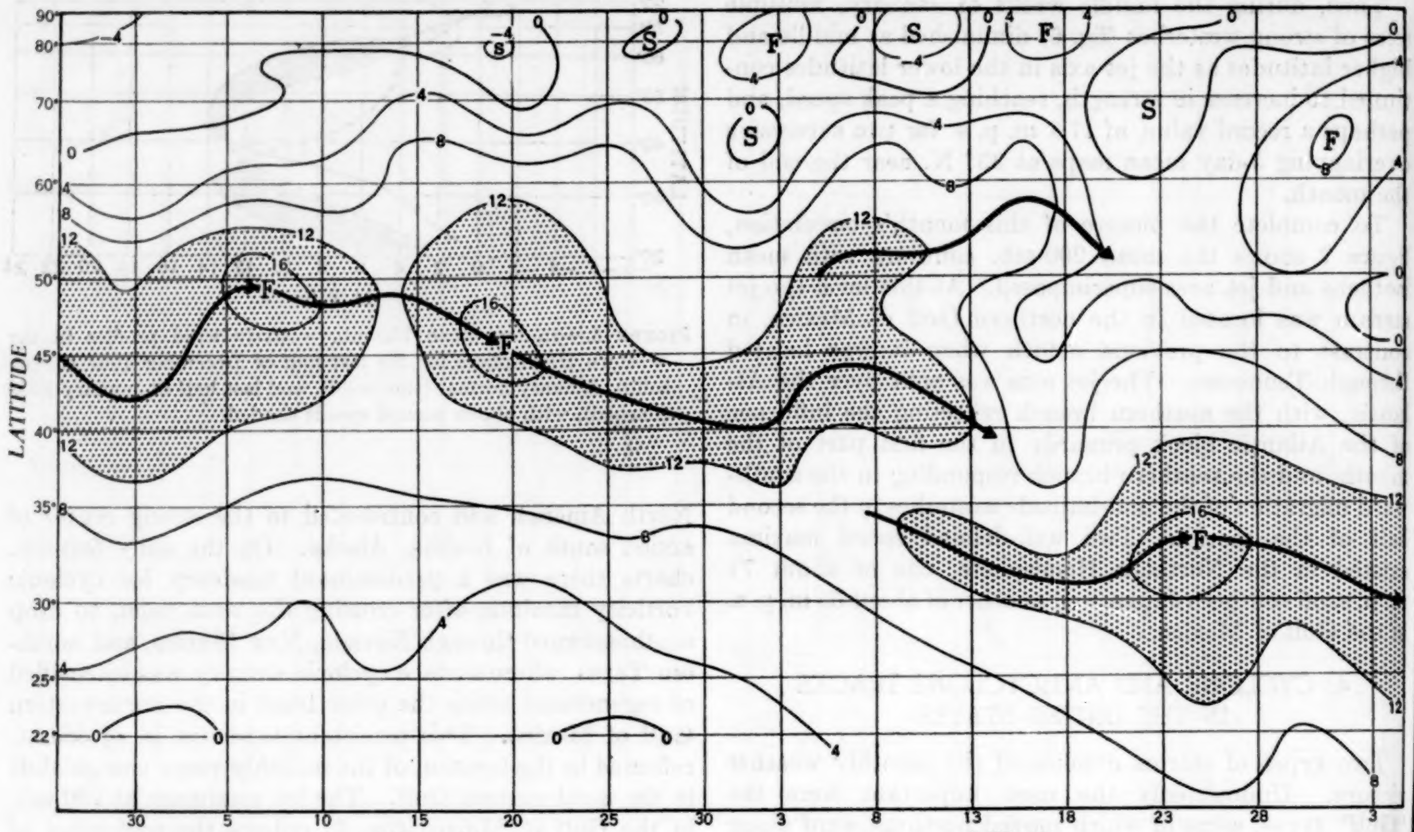

FIGURE 6.—Time-latitude section of 700-mb. 5-day mean wind speeds (m. p. s.) for December 1957 and January 1958, averaged over 5° of latitude around Western Hemisphere. The 5-day averages were computed three times weekly and plotted at middle day of period. Axes of maximum westerlies are indicated by heavy solid lines, and speeds over 12 m. p. s. are shaded. F indicates high speed centers, and S, slow centers.

after having spread heavy precipitation throughout the entire eastern United States. During this period a polar High moved eastward across southern Canada instead of along the more normal path southeastward across the middle Mississippi Valley. Another high center located near northwestern Hudson Bay on the 16th, intensified rapidly and drifted northeastward into Baffin Bay. It reached the record intensity of 1066 mb. on the 18th, eclipsing the previous record high of 1064 mb. in this area in January 1956 [1]. This High stalled the Texas storm near New England, resulting in almost continuous precipitation in that area until the 19th. (Another article in this issue, by Martin and Bucci, gives a detailed comparison of these two storms.) Persistent northerly surface flow due to the stationary storm in the Northeast again produced below normal temperatures east of the Ohio and lower Mississippi Valleys, with maximum negative temperature departures in the extreme Southeast for the third consecutive week. In Maine, temperatures were much above normal during this week due to the persistent easterly flow of mild maritime air resulting from the stalled Low and the block over Labrador.

In the latter part of this week a frontal system crossed the western United States into the central Plains and developed another low-latitude disturbance over the Southwest, causing beneficial snows in the western portions of the central and lower Great Plains on the 18th and 19th, where some sections had been without precipitation since mid-November. This system brought heavy precipitation to practically the entire eastern United States from the 20th to the 22d and one of the heaviest snows of record at Kansas City and Columbia, Mo.

This was quickly followed by another storm emerging from northeastern New Mexico into the northern Gulf on the 24th, reaching southern New England on the 26th, and producing 1 to 2 inches of rain along the Gulf and Atlantic Coasts. This cyclone was also blocked near New England, similar to the storm of mid-month, by a High moving across southern Canada to Labrador. This system caused a damaging tornado at Cochran, Ga., on the 24th, and snow and sleet as far south as northern Louisiana and Mississippi.

On the last days of the month a storm was moving eastward through Oklahoma producing heavy snow in northeastern Missouri and Illinois.

6. TEMPERATURE AND PRECIPITATION

Some of the salient abnormalities of the weather have already been discussed in connection with the circulation anomalies. In general the temperature regime over

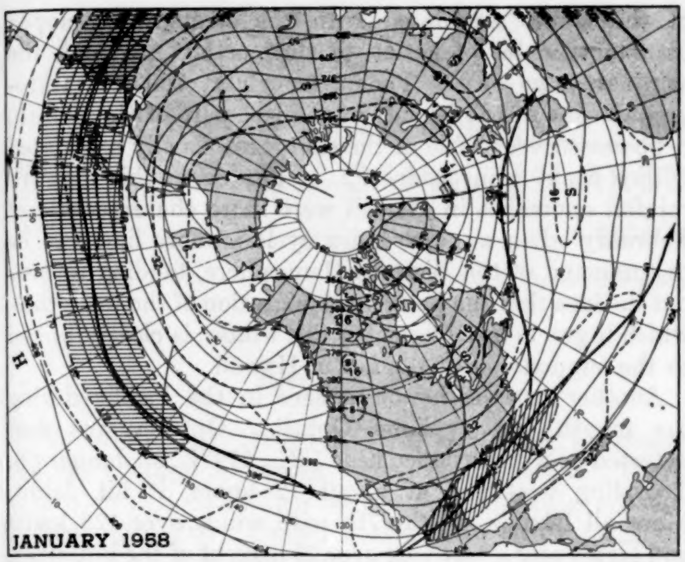

Figure 7.—Monthly mean 200-mb. contours (solid, hundreds of feet) and isotachs (dashed, m. p. s.) for January 1958. Heavy solid arrows indicate average position of the jet stream. (Hatching in areas with over 48 m. p. s.).

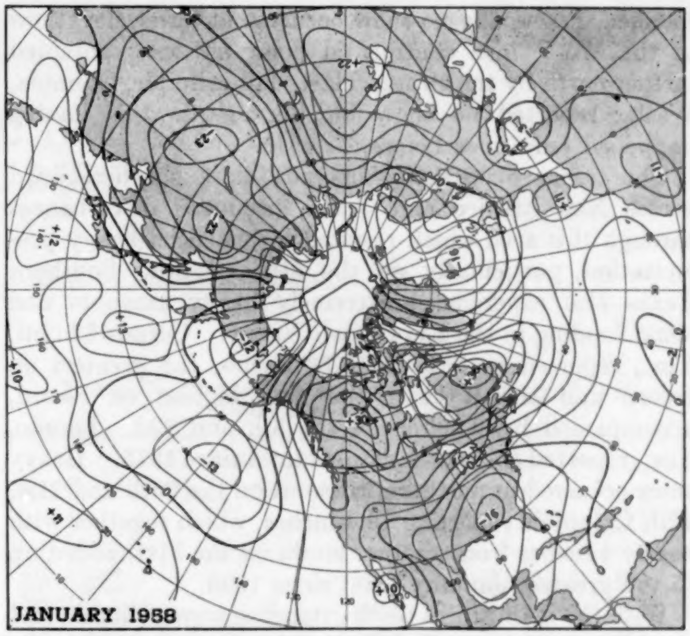

Figure 8.—Monthly mean departure from normal of 1000–700-mb. thicknesses for January 1958, drawn at 50-foot intervals, with centers labeled in tens of feet. Heavy solid lines are zero departures.

North America is well illustrated by the departures from normal of the monthly mean 1000–700-mb. thicknesses (fig. 8). Canada was completely dominated by above-normal thicknesses, with two distinct centers in the over-all pattern. The western center, averaging about 16F.° above normal, was sustained by the advection of mild maritime Pacific air in the stronger than normal southwesterly flow between the great planetary trough south of Kodiak, Alaska, and the ridge in western North America which persisted throughout the month.

The above-normal thickness center in Laborador, also averaging about 16F.° above normal for the month, developed and was sustained mostly in the second half of the month by advection of mild Atlantic maritime air by the strong southerly and southeasterly flow which developed in conjunction with the strong blocking in the Davis Strait (fig. 4B).

Although the thickness departure isolines in figure 8 might suggest that the thermal field was reversed in the northern United States and southern Canada, only in eastern Canada in the latter half of the month were the departures large enough to reverse the actual thermal field. For example, in the 5-day period January 18–22, the actual thickness at Frobisher, N. W. T., was greater than in Maine, and the mean thickness lines (not shown) were oriented from the Gulf of St. Lawrence northwestward, with colder air to the southwest. This helped to produce the abnormal retrogressive steering of surface Lows from the Maritime Provinces northwestward toward northern Hudson Bay (Chart X).

The temperature regime over the United States conformed closely to the height and thickness departures. Temperatures in the Southeast averaged 6 to 8 F.° below normal for the month (Chart I–B), with shorter averages, particularly for the second week, even more extreme. Most places in the extreme Southeast experienced their second coldest January of record (exceeded only in 1940). Elsewhere the Southeast had its coldest January in 10 years.

New England temperatures ranged from near normal in the south to 10° above normal in Maine due to the prevailing maritime trajectory, with the "January Thaw" arriving in most places in southern New England on schedule. The upper Mississippi Valley, Northern Plains, northern Rockies, and much of the western United States also experienced abnormally warm weather due to the persistent transport of mild Pacific maritime air resulting from the tenacious planetary wave pattern in western North America and the eastern Pacific, together with foehn effects in the lee of the northern Rockies. Marquette, Mich., reported the warmest January since 1934; Rapid City, S. Dak., was above normal on 30 days of the month; Havre, Mont., had no temperatures below zero and only one warmer January since 1879; and Los Angeles, Calif., had the warmest January on record.

Precipitation amounts (Chart III–B) were also spectacular in certain areas due to the persistent anomalies in the circulation. Above normal amounts fell in three principal areas: most of the Atlantic seaboard, the Southern and Central Plains, and most of the west coast. The eastern edge of the above normal precipitation in the West conformed closely to the position of the persistent planetary ridge line aloft, which terminated the moist ascending southwesterly currents from the eastern

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Pacific. Practically no snow occurred at lower elevations in the West since continental polar air was contained farther north by the strong ridge. For example, Olympia, Wash., reported no snow for the month, this having happened only once before, in 1945.

The tracks of vorticity maxima aloft, southeastward across New Mexico and Texas, produced disturbances through this area which resulted in beneficial heavy precipitation, particularly on the 5th and 6th. Southern Texas also experienced extremely heavy amounts and some flooding in the Rio Grande Valley. Corpus Christi, Tex., reported a total of 10.78 inches, the greatest on record and almost twice the next highest on record, accompanied by flooding on the 5th and 23d. Laredo, Tex., reported the second largest total since 1875. Heavy snow occurred in northern Missouri on the 20th and 21st, with Columbia reporting 10.4 inches, which together with nearly 4 inches from another storm on the 31st, added up to the greatest January total since 1906.

With the main storm track extending across Florida and northeastward just off the Atlantic coast, and with the jet stream and its associated speed maximum in the Gulf of Mexico, the whole Atlantic seaboard was vulnerable to heavy precipitation. Southern Florida had one of the wettest Januarys on record, and Charleston, S. C., reported the most rainfall of any January since 1929. Relatively lighter amounts occurred in the middle portion of the Atlantic seaboard, probably due to the coastal indentation north of Hatteras providing a measure of remoteness from the path of migratory cyclones.

Stagnation of storms off New England, due to the blocking to the north, resulted in record or near record amounts of precipitation at many stations; e. g., Burlington, Vt., reported 33.7 inches of snow, the most on record for any month. Boston's 9.54 inches of precipitation proved to be the greatest January amount in 88 years of record, producing some flooding of rivers. In addition Boston recorded 17 days with precipitation, 1 day less than the record.

7. THE TROPICAL PACIFIC

All Hawaiian stations reported precipitation below normal, probably due in part to the weakness or absence of the trades associated with the strongly developed planetary center of action to the north. Although the southern edge of the Pacific trough at 700 mb. retrograded west of the islands in the last part of the month, it possessed very little of the sharpness usually associated with a heavy rain producer. In any event this month's rainfall anomalies in Hawaii were more characteristic of February when a major index cycle is more likely to be in command of the Western Hemisphere, tending to bear out the idea that the southward migration of the westerlies, which began in late December this winter, is closely linked to the suppression of rainfall in Hawaii.

Another noteworthy occurrence in the Pacific during the month was typhoon Ophelia. Ophelia was first detected as a disturbance near 5°N., 174°E. on January 6. Traveling westward at about 12 knots, it hit Jaluit, Marshall Islands, on the 7th with winds over 125 knots and high water destroying over 95 percent of the buildings. Sixteen persons were reported dead or missing. Continuing westward at about 12 knots, it passed within 20 miles of Truk, Eastern Caroline Islands, on the 10th and 11th, yielding 3.42 inches of rain. Veering more to the northwest, it slowly made its way in the direction of the Philippine Islands but sharply recurved on the 17th at about 13°N., a few hundred miles east of the islands, where it dissipated. Figure 1 shows the extensive high pressure ridge in the western Pacific with its axis along the 18th parallel, which prevented the typhoon from penetrating into the westerlies.

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ABNORMALLY MILD TEMPERATURES IN THE CANADIAN ARCTIC DURING JANUARY 1958*

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1. INTRODUCTION

The winter of 1957–1958 has produced abnormal weather conditions in many parts of the world. There have been periods of unusually stormy and cold weather in the southeastern United States, and severe blizzards in the Great Lakes area of Canada and the United States as well as in the United Kingdom. On the other hand, record maximum temperatures for January have been established both in western and northeastern Canada. The most significant of these record high temperatures occurred in the northern Arctic islands.

Based on data contained in the regular coded surface synoptic weather reports and unchecked January monthly summaries, the highest temperature reported in the Arctic this January was 40° F. above zero at Arctic Bay on January 22 and 23. This temperature and the values of 32° F. at Alert and 30° F. at Eureka, both on January 24, are the only occurrences of near or above freezing temperatures ever reported in January from the Canadian Arctic north of latitude 70° N. In fact, these January 1958 maximum temperatures are higher than any temperature ever recorded previously during the 6 months from November to April inclusive.

2. MAXIMUM TEMPERATURES

Isotherms of extreme highest recorded temperatures for January are shown for the Canadian Arctic in figure 1. The solid isotherms represent the extreme maximum temperatures up to and including January 1958 while the dashed lines indicate the isotherms as they were previous to January 1958. It can be readily seen from this map that exceptionally mild weather was experienced in Baffin and Ellesmere Islands.

The period of record at each station is shown in table 1 which gives in tabular form the values mapped in figure 1. At Arctic Bay from 1938 to 1957 the highest January temperature recorded was 28° F. in 1945, while in this most recent January, 40° F. is shown as the maximum temperature on both January 22 and 23. Over a slightly shorter period of record at Frobisher from 1942 to 1957, the highest January temperature was 30°, while on January 21 this year a value of 39° was recorded. In southern Baffin Island the record high temperature of 48° recorded at Pangnirtung in 1931 was probably not

TABLE 1.—*Extreme highest recorded temperatures (F.°) in Northeastern Arctic Canada*

Code	Name	Previous Januarys			January 1958	
		Period	Maximum	Year	Maximum	Date
LT	Alert	1950–57	17	1957	32	24
EU	Eureka	1947–57	24	1952	30	24
IC	Isachsen	1948–57	20	1952	25	23
MD	Mould Bay	1948–57	15	1957	6	22
RB	Resolute	1947–57	22	1952	23	23
CR	Craig Harbour	1934–40	20	1940	-----	-----
DH	Dundas Harbour	{1931–33, 1946–48}	26	1947	-----	-----
AB	Arctic Bay	1938–57	28	1945	40	22, 23
PI	Pond Inlet	{1922–27, 1931–51}	20	1945	-----	-----
CY	Clyde	1943–57	31	1955	29	24
RO	Fort Ross	1938–48	19	1940	-----	-----
NC	Spence Bay	1951–57 B	8	1952	29	22
UX	Hall Lake	1956–57	12	1956	34	21
PO	Padloping Island	1941–56	31	1945	-----	-----
PU	Pangnirtung	1931–40	48	1931	-----	-----
FB	Frobisher	1942–57	30	1955	39	21
RE	Resolution Island	1929–57	34	1940	35	21

B=Broken period.

TABLE 2.—*Winter season monthly values of extreme highest recorded temperature (F.°) at representative stations*

Code	Name	Period	Period of record to 1957								January 1958
			Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	
LT	Alert	1950–57	33	31	13	17	25	28	30	47	32
EU	Eureka	1947–57	39	29	11	24	6	8	26	42	30
IC	Isachsen	1948–57	28	25	11	20	−8	17	35	36	25
RB	Resolute	1947–57	32	27	17	22	6	20	30	40	23
AB	Arctic Bay	1937–57	44	36	34	28	36	34	36	51	40
FB	Frobisher	1942–57	45	42	33	30	27	39	41	56	39
RE	Resolution Island	1929–57	45	39	35	34	34	37	39	45	35
KL	Knob Lake	1949–57	62	46	41	31	39	49	47	83	40
YR	Goose Bay	1942–57	73	60	53	42	51	54	62	89	46

exceeded this year (1958) although the lack of an observing station at this location at present leaves the matter in doubt.

Mean temperatures during January 1958 for northern Baffin Island and Ellesmere Island were in the neighborhood of −15° to −20° F. Farther west, stations on the smaller islands had mean temperatures lower than −20°. This January was the warmest January on record at five Arctic weather stations—Alert, Eureka, Resolute, Isachsen, and Mould Bay. The departure from normal in northern Ellsmere Island was 16 F.° which is the greatest positive temperature anomaly ever observed in January in far northern Canada. There have been, however, several warmer Januarys on Baffin Island. At Arctic

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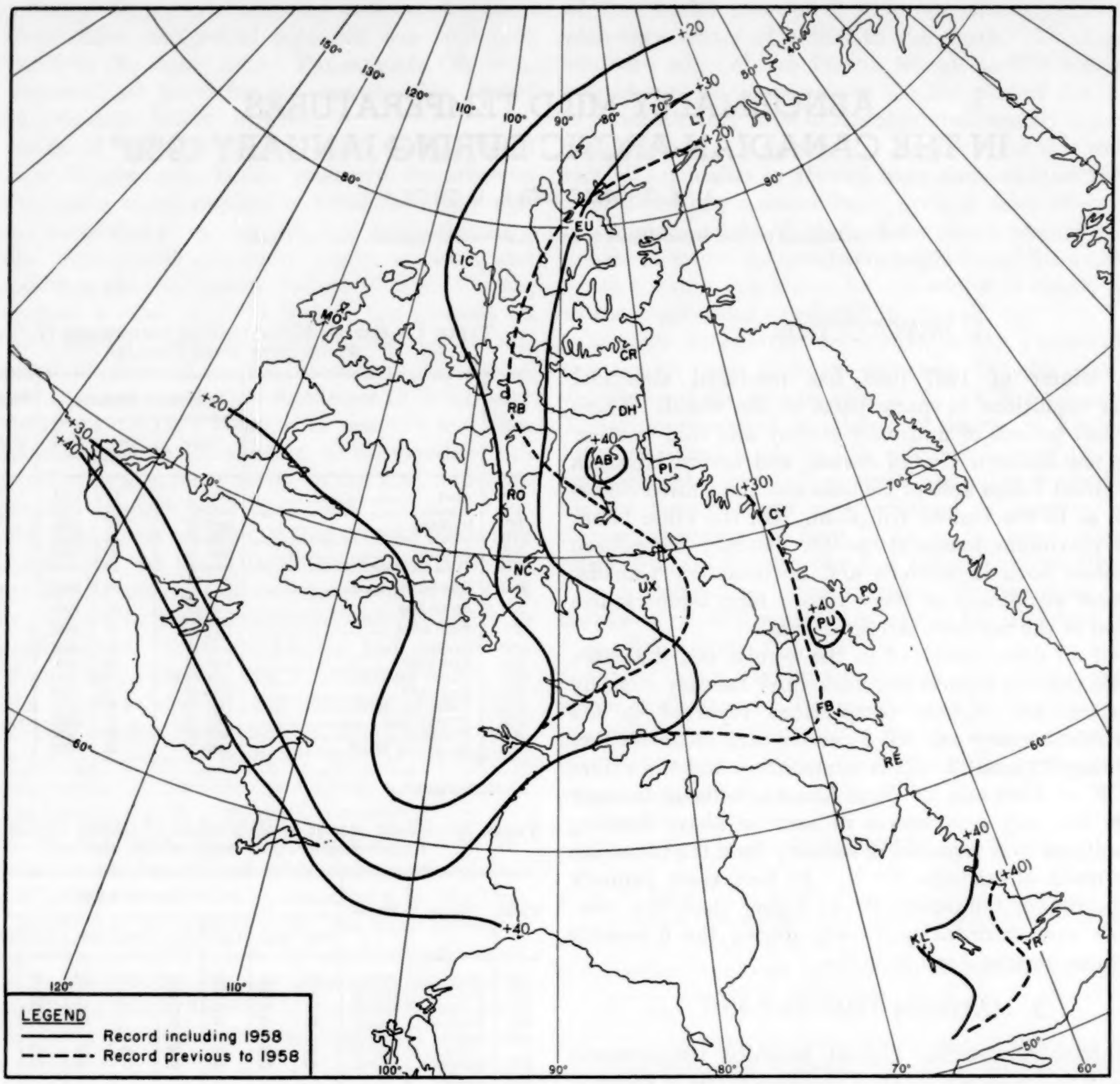

FIGURE 1.—Extreme highest recorded temperatures in January.

Bay the mean temperature was −17° compared to a normal value of −22° while the warmest January on record, 1940, had a mean temperature of −1°. Along the coast at Clyde, mean temperatures over the month of January 1958 were normal.

With the exception of a 3-year period from 1928–1930, temperature records are available since 1921 from one or more stations in the Canadian Arctic north of latitude 70° N. During this time, above freezing temperatures in January have never been reported previous to 1958. The years with extremely warm periods in January have been 1940, 1945, and 1952, when maximum temperatures in excess of 20° were reported from stations such as Arctic Bay, Resolute, and Eureka. Temperatures in excess of 20° have been reported slightly more frequently at such coastal stations as Clyde, Pond Inlet, and Craig Harbour.

An interesting situation was observed during the month at Clyde where the temperature remained low after the warm front had passed. On January 23 when the other Baffin Island stations had temperatures near or in excess of 32° F. the Clyde temperature remained below zero. This was probably because of the topography at Clyde which acted to trap the cold air at the surface while the warmer Maritime Arctic air passed over the top of the fjord in which the station is located. The cold air eventually moved out and on January 24 Clyde reported a maximum temperature of 29°.

There were at least 7 or 8 stations in Ungava and Labrador where January high temperature records were broken. The most significant instances were temperatures of 46° at Goose Bay and 39° at Frobisher where the previous maxima had been 42° and 34° respectively. Also, at Resolution Island, where a relatively long record dates back to 1929, a new January high of 35° was estab-

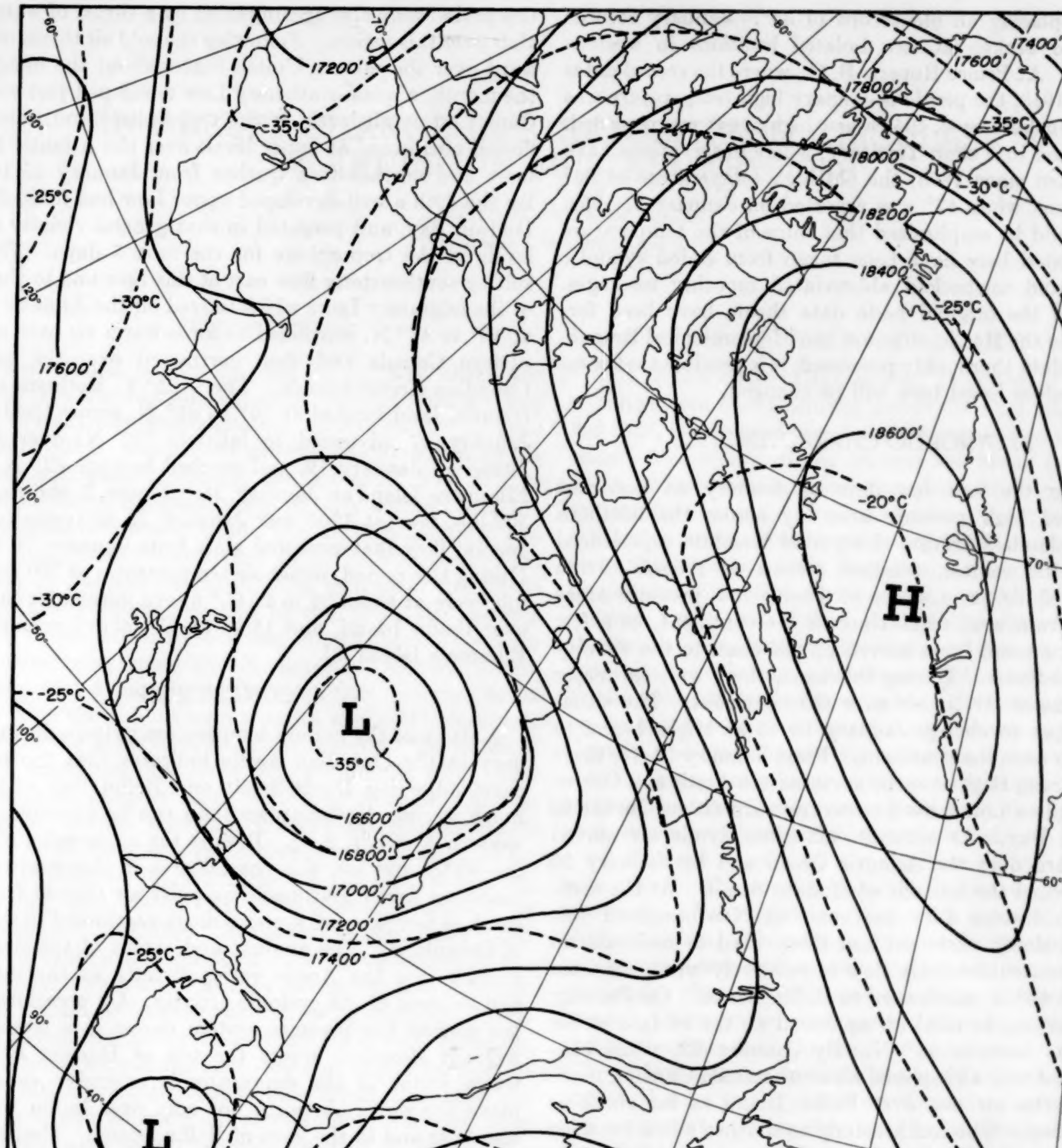

FIGURE 2.—500-mb. contours and isotherms, 1200 GMT, January 22, 1958.

lished replacing an old record of 34°. January records were also broken at two isolated locations in western Canada. At Prince Rupert, B. C., where the record dates back to 1913, the previous January high temperature was 62° but on January 6, 1958 a new high of 64° was recorded. At Dawson in Yukon Territory, where observations have been taken since 1898, the old high temperature of 36° was broken when 47° was reported on January 7, 1958.

It should be emphasized that some of the temperature values listed here have been taken from coded synoptic reports and unchecked abbreviated monthly messages. When all the original basic data sheets have been forwarded to the Headquarters of the Meteorological Branch, and the data thoroughly processed, it is possible that some of the values listed here will be changed.

3. SYNOPTIC CHARACTERISTICS

During the first few days of January an east-west orientated high pressure area lay across the northern Arctic islands. During this period Resolute experienced one of the coldest, windiest storms on record. From January 7 to 12, a series of Pacific low pressure areas moved from west to east across the continent, as a few Atlantic coastal Lows moved up the coast to the vicinity of Davis Strait. During this time, there were no major outbreaks of Arctic air over the continent. The situation began to change January 13–15 as Highs began to build up over the continent. From January 16–19, there was a strong High over the northeastern Arctic and Greenland while a Low moved northwestward across Quebec to Hudson Bay. As a result, maritime Arctic air moved northward over the Atlantic Ocean and by January 20 had reached the latitude of Hudson Strait. At the same time an intense Low southeast of Newfoundland was drifting slowly eastward and this served to maintain an unbroken southeasterly flow of relatively warm air from latitude 40° N. northward to Baffin Island. On January 21, maritime tropical air appeared at the surface as far north as latitude 48° N. By January 22, while continental Arctic air covered all southeastern Canada, maritime Arctic air was over Baffin Island as far north as Arctic Bay. The southeasterly to southerly flow between the High over Greenland and the Low persisting in the vicinity of Hudson Bay continued to sweep the maritime Arctic air northward, and this air reached the mid-part of Ellesmere Island on January 23. On this date most stations on Baffin Island reported temperatures of 32° F. or higher. On January 24 continental Arctic air swept eastward over Baffin Island, and by the 25th had reached Ellesmere Island, bringing to an end this regime of abnormally mild weather in the Canadian Arctic. Just before the invasion of this colder air, Alert, near the north shore of Ellesmere Island, reported a 38° F. reading at 2,000 ft. above the surface.

As might be expected, the abnormally warm temperatures at the surface level in northeastern Canada during this period were also accompanied by a thrust of warm air aloft over this region. Following the cold air thrust southward over the eastern United States about the middle of the month, a quasi-stationary Low developed just west of Cape Cod at all levels in the troposphere and warm air flowed northward at upper levels over the Atlantic Provinces and northeastern Quebec from January 16 to 19. By the 19th a well-developed upper Low had formed over Hudson Bay, and persisted in that general vicinity at all levels in the troposphere for the next 5 days. The resulting southeasterly flow east of this Low and to the east of the migratory Lows which moved off the Atlantic coast at 40° to 45° N. continued to force warm air over northeastern Canada and then northward over the eastern Canadian Arctic Islands. The −25° C. isotherm at the 500-mb. level located at 50° to 55° N. across Quebec on January 17 advanced to latitude 70° N. over Baffin Island by January 19, and reached latitude 82° N. over Ellesmere Island on January 24. Figure 2, showing the 500-mb. flow at 1200 GMT January 22, is typical of the general flow that persisted aloft from January 19 to 24. During this period, upper-air temperatures at 700 and 500 mb. were at times 20 to 25 C.° above normal for January over Baffin Island, and 15 to 20 C.° above normal over Ellesmere Island [3].

4. CONCLUSION

A study of the normal temperature pattern during January in the Canadian Arctic indicates that the eastern coast adjoining Davis Strait and Baffin Bay often has much warmer temperatures than can be expected in the western Arctic [1, 2, 4]. Besides the moderating effect of the water and ice, this coastal area is also subjected to maritime Arctic and maritime polar air thrusts from the south as low pressure systems move northward to the west of Greenland. The amount and extent of the warm air moving into the Arctic varies directly as the intensity and location of the cyclonic activity. On occasions when the surface low pressure system moves in a west-northwesterly direction across the top of Hudson Bay, the warm sector of the depression is extensive enough to move the warm air north not only over Baffin Bay but also over and to the west of Baffin Island. That is what occurred on January 19–24, 1958 and as a result the northeastern Arctic experienced some abnormally high wintertime temperatures.

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AN ABRUPT CHANGE IN STRATOSPHERIC CIRCULATION BEGINNING IN MID-JANUARY 1958

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1. INTRODUCTION

Observational studies referred to by Kochanski and Wasko [1] show that winds at the 25-mb. constant pressure surface (altitude approximately 82,000 ft. or 25 km.) north of 30° N. latitude are generally westerly during the winter and easterly during the summer. The change from the summer regime begins in early autumn with the appearance of westerly winds, first in high latitudes and then progressively southward to the subtropics. With cooling at the pole during the first months of the Arctic night, low pressure develops in the polar stratosphere and is accompanied by a strong stratospheric jet stream of westerlies at about 70° N. latitude. Generally during the first 3 months of the year, low pressure splits away from the pole to form troughs, and anticyclogenesis occurs in the areas between these troughs. These latter events sometimes take place with a dramatic effect upon stratospheric temperatures. Scherhag [2] was the first to report an abrupt warming under these circumstances. In years following, this phenomenon was investigated by Scrase [3], Wexler [4], Warnecke [5], Lee and Godson [6], and Craig and Hering [7]. Wexler and Moreland [8] ascribed the wintertime "breakdown of essentially zonal flow into large-scale meridional flow" to instability of the stratospheric west winds as they strengthen into a powerful jet stream to balance the increasing meridional temperature gradient during the first months of polar darkness.

A series of six contour and isotherm charts for the 25-mb. surface is presented here to show the breakdown of the polar Low and concomitant stratospheric warming as it transpired in the period following January 17, 1958. The maps showing the significant stages of this breakdown were selected from daily charts for the 50-mb. and 25-mb. surfaces, plotted by the National Weather Analysis Center since June 1957.

Whenever possible the contour analyses of the charts were based on observed data. Unfortunately large areas of sparse data are quite common above the 50-mb. level. This is true even over areas with an adequate station network, for high winds cause balloons to drift out of range and extremely low temperatures lead to premature balloon bursts. In areas where data were scanty, 25-mb. contours were constructed by addition of the 50-mb. contours and the mean thickness of the 50- to 25-mb. layer.

2. DISCUSSION OF CHARTS

The 25-mb. chart of January 17, 1958 (fig. 1) is typical of the pattern that gradually developed after late September 1957, interrupted occasionally by short-period fluctuations. The dominant feature was the broad flow of the Arctic stratospheric jet stream coupled with extremely cold temperatures over the Greenland area. From geostrophic estimation, wind speeds approaching 200 kt. were probable along the jet axis. Over the Aleutians, abnormally high temperatures, associated with a large area of high pressure, became firmly established during the latter part of December. By January 17, some of the warmth had penetrated slowly across Alaska and into central Canada. To show the contrasting stratospheric airmasses, soundings for Bethel, Alaska, to represent the warm air, and Eureka, Canada, the cold air, are reproduced in figure 7. There is a strong similarity between the two soundings below the level of the tropopause, but upward from that point the airmass contrast becomes very pronounced.

By January 24 (fig. 2) a major change had begun to take place. The most noticeable feature was eastward movement of the Aleutian High to a position over the Gulf of Alaska. Cold air persisted over Greenland with an extension southwestward. Colder temperatures than previously observed appeared over the eastern United States. Over central and eastern Canada, the contour and temperature fields had moved into phase.

Four days later on January 28 (fig. 3) a lobe of the cold pocket, that on earlier charts was located over Greenland, had begun to penetrate the Great Lakes area. Examination of intervening charts reveals that horizontal advection does not explain the movement of this cold air pocket. If temperature changes are adiabatic, upwelling of isentropic surfaces is occurring. This is a counterpart phenomenon to the subsidence that permitted temperature rises in spite of very strong horizontal cold air advection over Newfoundland in January 1957, as described by Craig and Hering [7]. South of Greenland, widely scattered reports give evidence of pronounced stratospheric warming, and the −39° C. isotherm can be drawn with reasonable confidence. This is in sharp contrast with temperatures of less than −70° C. that were in that area just 4 days previously. A small area of warming was also developing off the southeastern coast

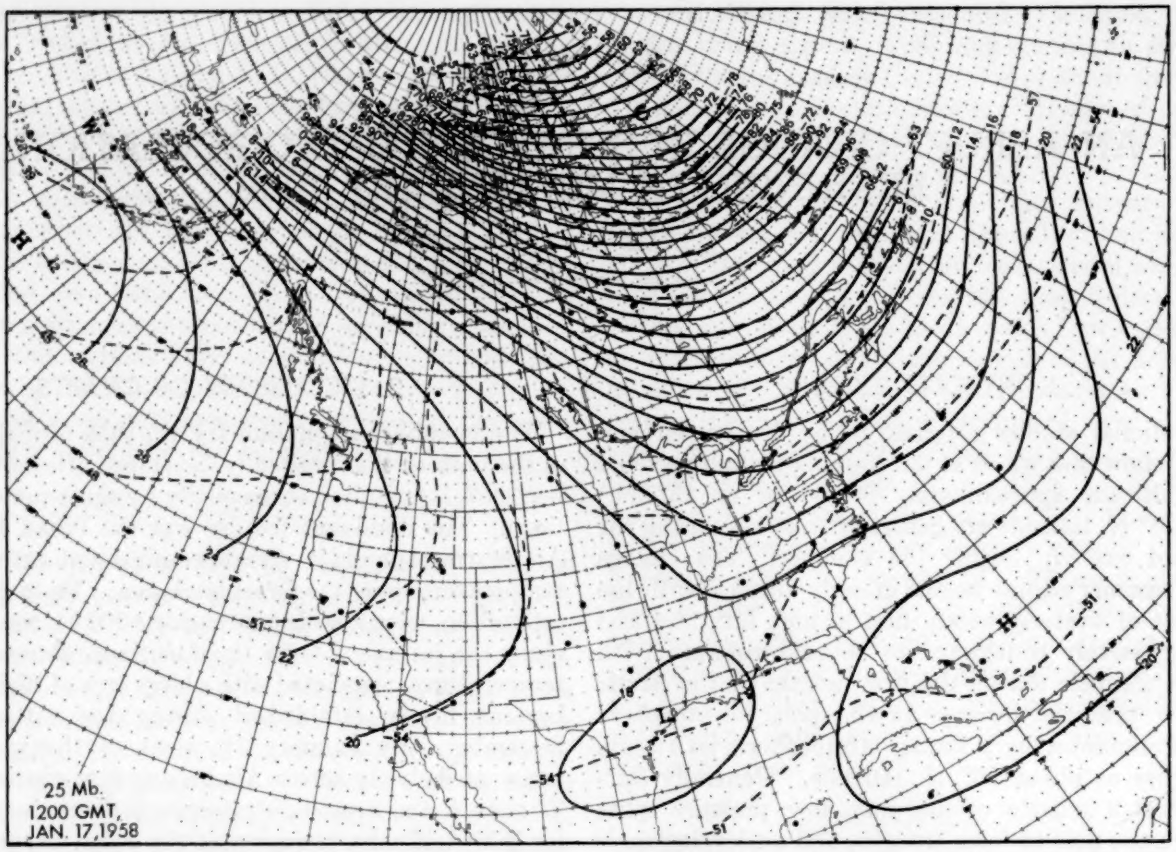

FIGURE 1.—25-mb. chart for 1200 GMT January 17, 1958. Isotherms (dashed lines) are in degrees Celsius. Contours (solid lines) are drawn for 200-ft. height intervals, and labeled in thousands and hundreds of feet. Dots represent stations where data were available at map time or within 12 hours.

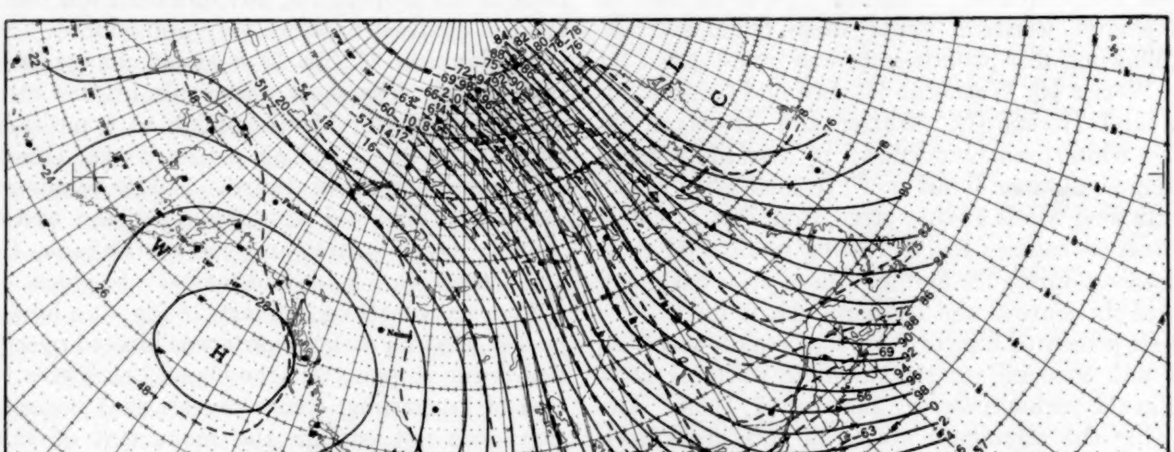

FIGURE 2.—25-mb. chart for January 24, 1958. Isopleths and dots as in figure 1.

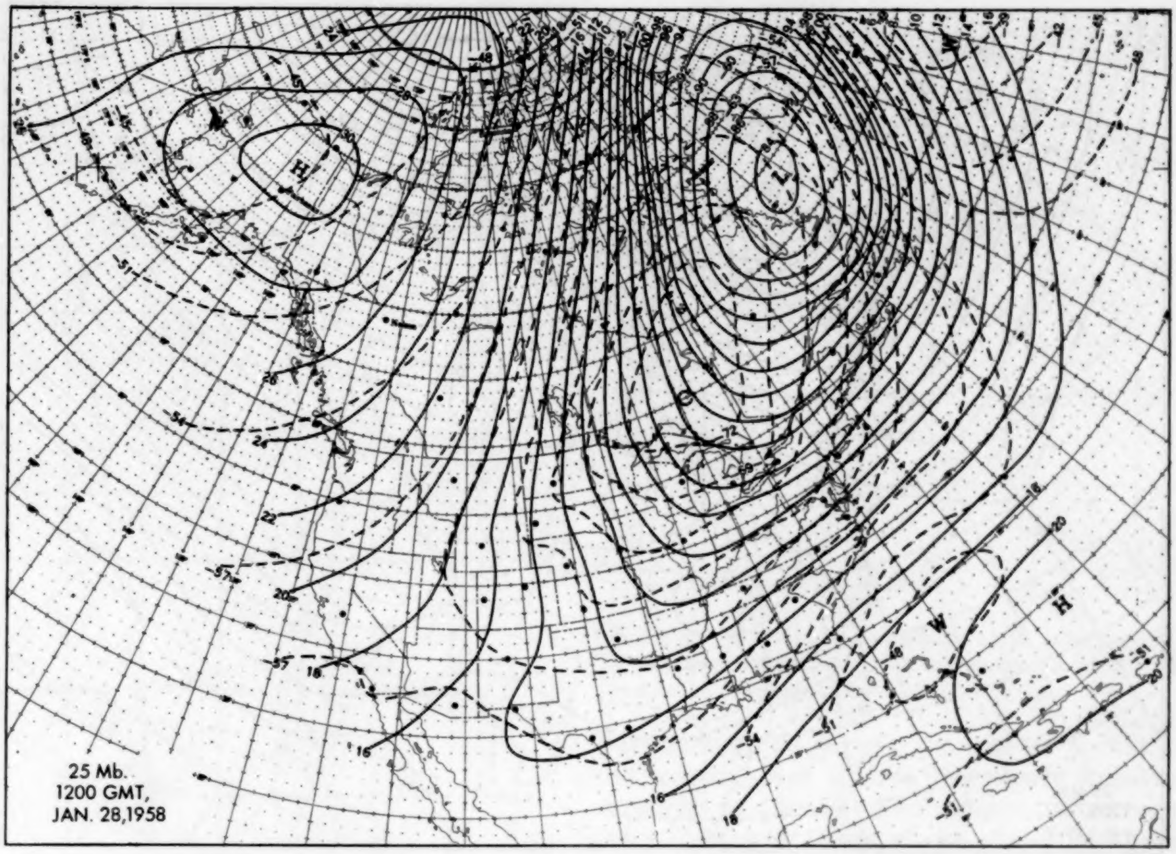

FIGURE 3.—25-mb. chart for January 28, 1958. Isopleths and dots as in figure 1.

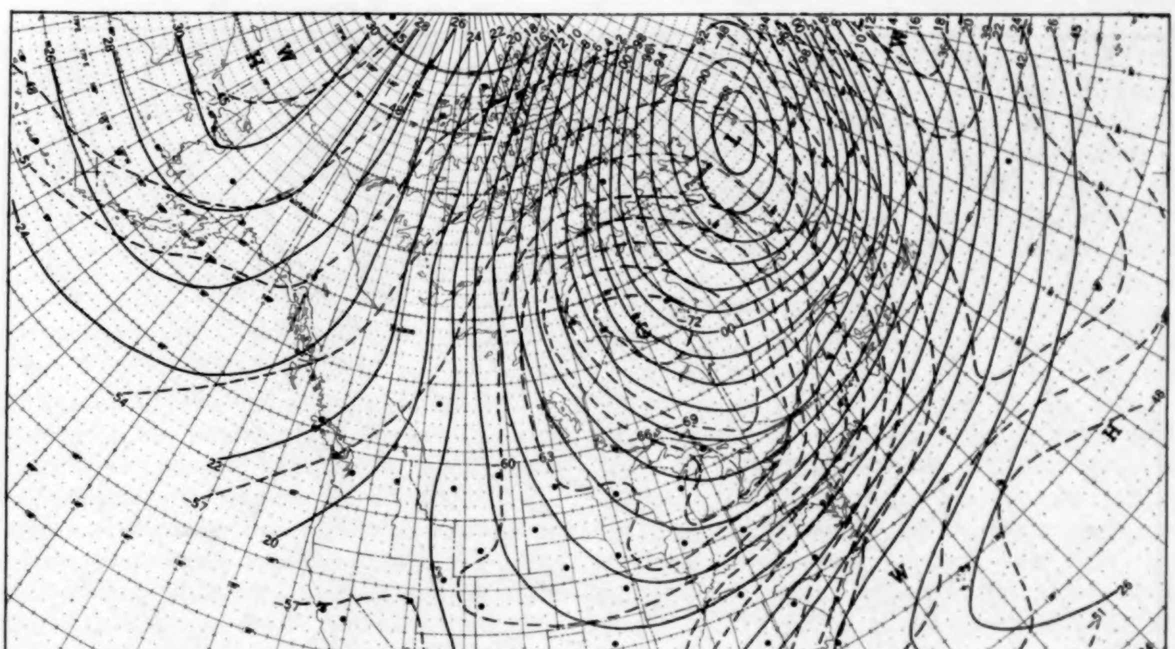

FIGURE 4.—25-mb. chart for January 29, 1958. Isopleths and dots as in figure 1.

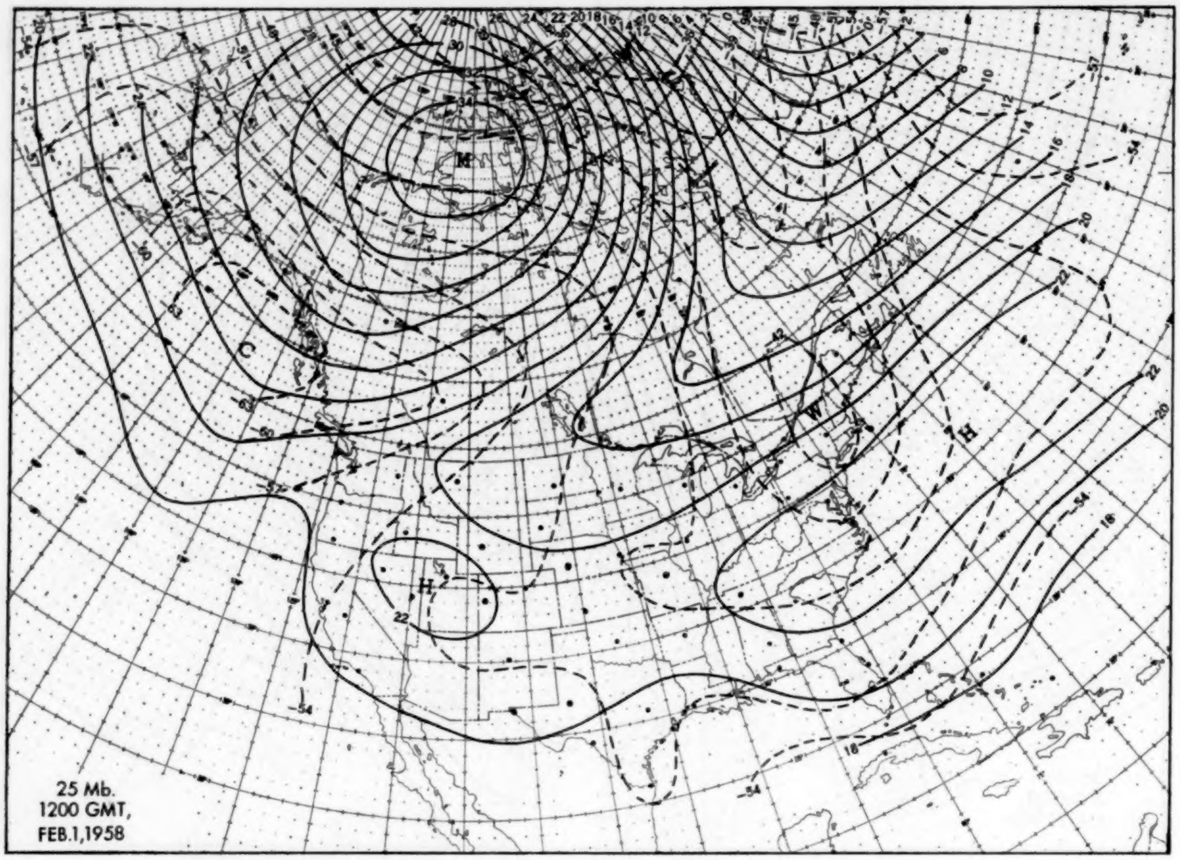

FIGURE 5.—25-mb. chart for February 1, 1958. Isopleths and dots as in figure 1.

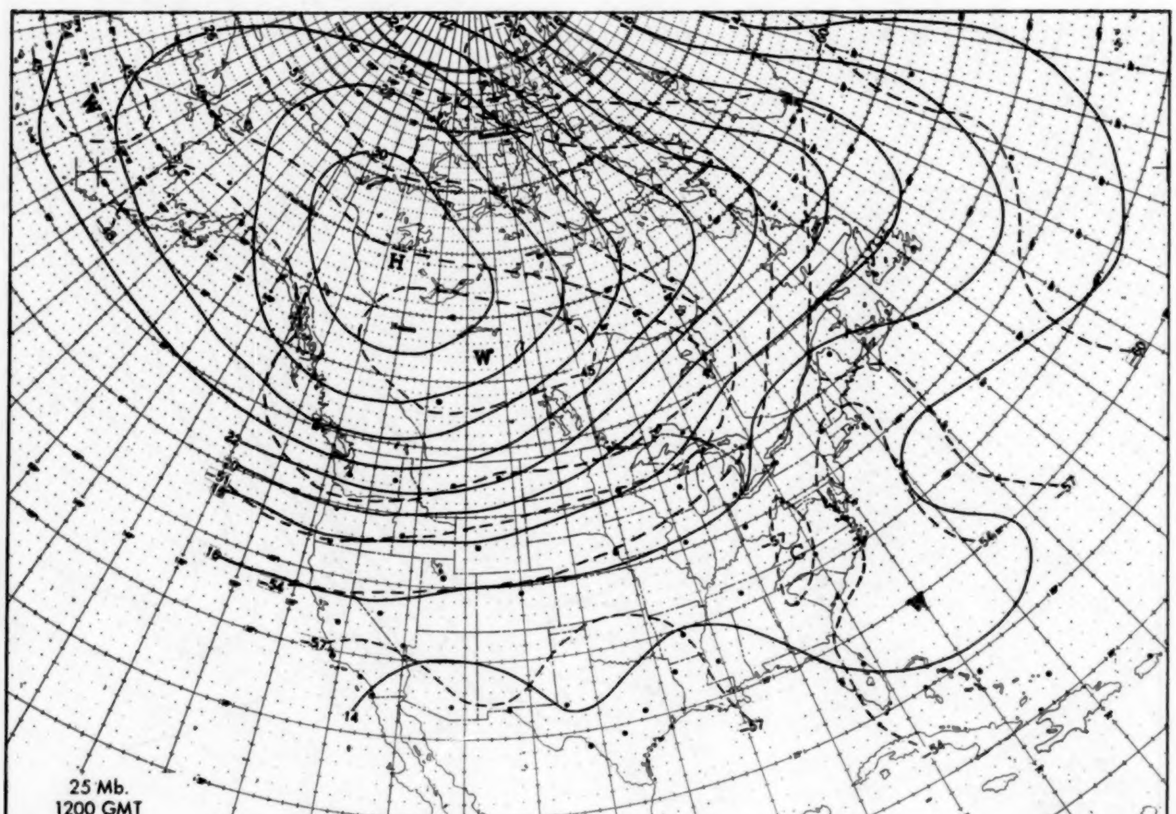

FIGURE 6.—25-mb. chart for February 5, 1958. Isopleths and dots as in figure 1.

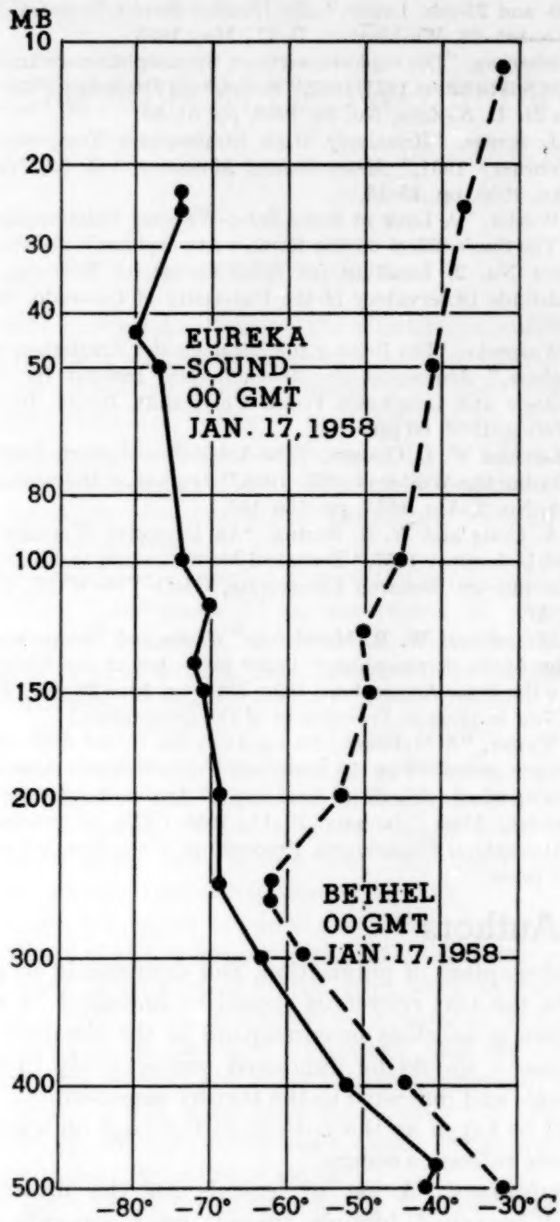

FIGURE 7.—Representative soundings taken in the warm and cold air at beginning of period.

of the United States. This warm area may already have been connected with the one south of Greenland, as it was in later charts, but scarcity of data over the Atlantic makes this uncertain. The association of stratospheric anticyclogenesis and warming with increase of total ozone content has been discussed in detail by Wexler [9]. The phenomenon is vividly illustrated in this case by total ozone content of 0.48 cm. and 0.49 cm. measured with the moon as light source at Alert and Resolute at about the time of figure 5. These provisional values, furnished through the courtesy of the Canadian Meteorological Service, represent increases of 0.25 cm. and 0.13 cm. from the immediately previous measurements made 4 weeks earlier. In the computations, use was made of the Vigroux absorption coefficients, which give ozone values about 36.5 percent larger than those given by the older Ny and Choong coefficients.

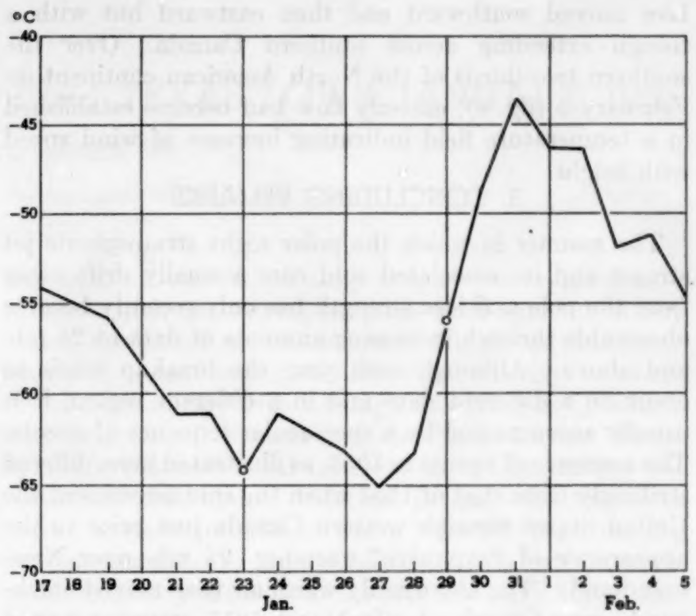

FIGURE 8.—1200 GMT temperatures at 25 mb. over Washington, D. C., from January 17, 1958 to February 5, 1958. Circles show estimated temperatures.

By January 29 (fig. 4), the cold air pocket at 25 mb. was centered over Hudson Bay slowly shrinking in size obviously as a result of subsidence. This conclusion is also indicated by 50-mb. charts where, on the same dates, the cold pocket was displaced to the northeast of its position at 25 mb. The lowest observed temperature was at least 6 C.° higher on January 29 than 5 days earlier. Stratospheric warming was well established (fig. 4) with a maximum somewhere east of Greenland and a warm tongue extending southwestward across Florida. The highest 25-mb. temperature noted from available data was −19° C. at Keflavik, Iceland (off the map). The warm air located just off the southeastern United States coast in figure 4 seems to have originated in the eastern Caribbean area. While moving northward, it had intensified and was about to merge with the main body of warm air to the north. At 1800 GMT of January 29, a temperature of −27° C. was observed at the 8-mb. surface (about 106,000 ft.) above Washington, D. C. during one of a series of special soundings released from Andrews Air Force Base. This observation is interesting because it indicates the degree of vertical slope of the warm air (at 13 mb., or about 96,000 ft., the temperature at both Hatteras and Washington was −44° C.). At 25 mb., the temperature at Washington (fig. 8) was already rising rapidly, from −65° C. on January 27 to −43° on January 31.

On the 25-mb. chart for February 1 (fig. 5), the cold pocket had drifted with some modification to the Gulf of Alaska. The 25-mb. contour and isotherm patterns were then in striking contrast to those of January 17. Warm air covered Greenland and extended southwestward into the eastern United States. In southern Alaska, temperatures were nearly 20 C.° lower than before. As the great anticyclone drifted into northwestern Canada, the polar

Low moved southward and then eastward but with a trough extending across southern Canada. Over the southern two-thirds of the North American continent on February 5 (fig. 6), easterly flow had become established in a temperature field indicating increase of wind speed with height.

3. CONCLUDING REMARKS

The manner in which the polar night stratospheric jet stream and its associated cold core annually drift away from the pole and lose strength has only recently become observable through increasing amounts of data at 25 mb. and above. Although each year the breakup tends to occur on a different date and in a different region, it is usually accompanied by a spectacular sequence of events. The sequence of events in 1958, as illustrated here, differed strikingly from that of 1957 when the cold air entered the United States through western Canada just prior to the appearance of "explosive" warming 25 mb. over Newfoundland. The abnormally warm air then moved northward across Greenland. In March 1956, warming moved into central and southern Canada from the Alaskan region but was not accompanied by a rapid breakdown of the circumpolar Low. Other cases studied, dating back to 1948, show a variety of breakup times and magnitudes of stratospheric warming.

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In *equations* conventional symbols in accordance with the American Standards Association Letter Symbols for Meteorology should be used. If equations are written into the manuscript in longhand, dubious-looking symbols should be identified with a penciled note.

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Illustrations. A list of legends for the illustrations should be typed (double spaced) on a separate sheet. Each illustration should be numbered in the margin or on the back outside the image area. To fit into the *Review* page, illustrations must take a reduction not to exceed 3½″ x 9″ (column size) or 7½″ x 9″ (page size). Map bases should show only political and continental boundaries and latitude and longitude lines, unless data are to be plotted, when station circles will also be needed. Usually the less unnecessary detail in the background the better will be the result from the standpoint of clear reproduction. Line drawings and graphs should also be uncluttered with fine background grids unless the graph demands very close reading. It is not necessary to submit finished drawings, as drafting work can be done at the time the paper is prepared for publication.

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A COMPARISON OF TWO REDEVELOPING TEXAS LOWS, JANUARY 1958

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1. INTRODUCTION

It is of interest to examine the similarities and differences in the cyclonic development of two storms that produced extensive rain or snow over the Southern and Eastern States, followed by freezes along the Gulf coast and over Florida during January 1958. This paper discusses the significant and noticeable features of the path, intensity, and weather of the storms through their cycle of development.

These two Lows appeared first in the zone outlined by Bowie and Weightman [1] as the region of origin for Texas type storms and met the other initial condition for Texas storms, namely: high pressure over the Eastern and Northwestern States. Texas storms follow a course toward the northeast or east over the Gulf States then northeastward along the Atlantic seaboard. Although figure 1 shows that the Lows of this study followed generally a normal track, a more critical look at the progress of these Lows shows that they posed entirely different problems for the forecaster.

To avoid subjective opinions about the relative depths or strengths of the storms, a quantitative measure is used to evaluate the strength of the two Lows as they are discussed here. The intensity is evaluated according to the number of standard deviations that the central pressure departs from the mean central pressure of many other Lows at the same latitude. James [2] proposed this as an appropriate statistical measure of the intensity of low and high pressure systems.

For convenience in distinguishing the two Lows in this paper, the earlier one that was first analyzed as a closed center on January 5 will be referred to as Storm A and the later one on January 12 as Storm B.

2. ANTECEDENT CONDITIONS

SURFACE

The two storms formed less than 150 miles apart. Storm A, which started to organize on January 5 about 120 miles south of Brownsville, Tex., would be classed as a center of weak intensity. Its central pressure of 1015 mb. was more than two standard deviations above the normal central pressure for Lows at that latitude. Storm B, which formed a week later, on January 12, about 40 miles northwest of Brownsville, would be classed as a Low of normal intensity. Its central pressure of 1007 mb. was within one standard deviation of normal.

As these storms took shape, certain features of the surface synoptic charts were strikingly similar. Storm A, the least intense in its early stages, was related to a 1044-mb. high cell moving southeastward over the eastern States, a 1042-mb. quasi-stationary High over the Northwest and to an active 980-mb. Low moving eastward across Hudson Bay with a trough extending to the southwest over central Canada and the Plains States. Storm B, which was of normal intensity in the beginning, was associated with almost identical pressure features with respect to position. There was a 1033-mb. High moving southeastward over the Eastern States, a 1026-mb. quasi-stationary High over the Northwestern States, and a 993-mb. Low crossing Hudson Bay. A significant point here is the fact that Storm B, though classed as more intense, was actually associated with less circulation and vorticity than Storm A, inasmuch as the accompanying

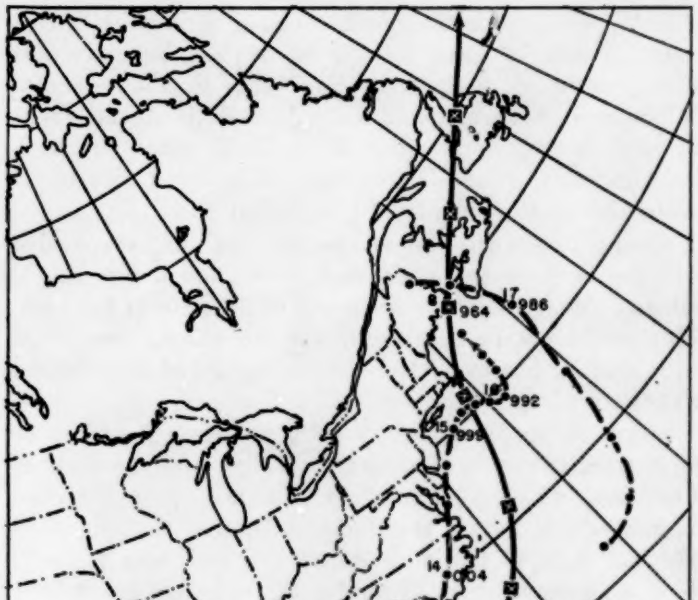

FIGURE 1.—Tracks of storm A, January 5–8, and storm B, January 12–16, 1958.

pressure systems were each roughly 10 mb. less intense than those with Storm A, and this is within the range of one standard deviation for all latitudes. This may be the first anticipation of modification in developments.

There were other dissimilar features. With the beginning of Storm A, a very intense quasi-stationary Low was in the Bering Sea and a second Low was centered about 10° of latitude south of ship station Papa (50° N., 145° W.), resulting in strong southerly flow over the eastern Pacific that maintained or had a tendency to increase the High over the Northwest. Associated with Storm B was a small, diminishing Low in the Pacific about 200 miles southwest of Annette Island with a frontal system extending to the south and a large deep Low dominant in the central North Pacific, about latitude 45° N., moving rapidly eastward. The southerly flow along the Pacific coast was not as strong in this situation with the main Low farther west and there was a tendency for the High over the Northwest to lessen. In the immediate vicinity of Storm B there was a weak Low and associated cold front to the northwest near Midland, Tex., while Storm A had strong northeasterly gradient to the west of it and a small ridge pushing down from the large High area farther to the north.

Both storms moved northeastward from the Texas coast. Storm A continued on to the east but Storm B swung more northerly over Louisiana, then east across Mississippi, Alabama, and Georgia. Both storms redeveloped as they approached the Atlantic coast, Storm A insignificantly near Jacksonville, Fla., and Storm B noticeably in the vicinity of Charleston, S. C.

As the two storms moved across the Gulf of Mexico and the Gulf States, they were accompanied by general rains and gusty winds to the north of the path of the Low. This would not be considered weather closely associated with fronts; however, the developments of the frontal waves were important.

Storm A produced heavy rains totalling 6 inches or more over the lower Rio Grande Valley and the Coastal Bend and widespread snow in amounts up to 7 inches over western Texas, the Panhandle, and eastern New Mexico. Corpus Christi had 7.72 inches of rain January 4–5, with considerable flooding and damage from high winds. Gusts as high as 70 knots were reported in the Corpus Christi area and the highest tide, 4.9 feet MLW, at Navigation Barge Bridge, was the highest since 1933. Heavy rains and gales accompanied Storm A near its center as it crossed the Gulf. Storm B caused moderate to heavy rains over southern Arkansas, Louisiana, and Mississippi, with gusty winds to about 35 knots. During the period prior to redevelopment, the storms caused widespread but generally light amounts of rain in the Southeastern States.

As Storms A and B moved eastward, the Hudson Bay Lows moved rapidly eastward with their troughs more or less stationary along the St. Lawrence River Valley. The Highs over the Eastern States traveled out over the Atlantic to a position directly ahead of the two storms. The High with Storm A moved on east and did not block the path of the storm. However, the High ahead of Storm B ridged back to the northwest to join a High moving down over central Canada. This was the beginning of blocking conditions, similar to those in the second case described by Austin [3], which made the major difference in later development of these two storms. The western High with Storm A remained stationary, but the western High with Storm B dissipated rapidly ahead of the frontolyzing occlusion that moved in from the Pacific. A ridge from the eastern Pacific High immediately built over the Northwest but it was not the same type of persistent High that remained with Storm A, and the remnants of the trough associated with the occlusion later became the flat, weak trough along the west Texas border.

UPPER AIR

The 500-mb. analyses which characterize the upper level conditions associated with the early development of these storms help point out similarities and differences. On January 5, there was a low center just east of Nantucket with a trough to the southwest roughly parallel to the Atlantic coast line. A vast High centered just northwest of San Francisco ridged to the north along the Pacific coast so that most of the United States was in a general northwesterly flow from the west coast ridge to the east coast trough with a small ridge over the east central United States and a minor trough from the Midwest to a low center 200 miles south of Phoenix, Ariz.

Along with Storm B there was a similar progression, a trough along the Atlantic coast, and a ridge over the Central States separated by a trough from a western ridge—but the magnitude of these features was noticeably different. The two ridges were nearly equal in intensity but the western ridge was over the Rocky Mountains and southern California with a trough off the Pacific coast. This shorter distance between ridges made the midwestern trough much sharper as it extended to a low center just north of Midland, Tex. The trough off the Atlantic coast with Storm B was orientated more north-south from a Low over Nova Scotia.

The Lows in the Southwest were the most closely associated upper-level features of these two storms. The Low over Midland, with Storm B, was well to the east of Storm A's Phoenix Low, and although the area of closed circulation about the Midland Low was smaller, its central height was about 150 feet lower than that of the Phoenix Low.

As Storm A moved eastward the trough over the Midwest moved to the western Great Lakes and a new Low formed over Lake Superior indicating deepening. The flow from this trough eastward became more westerly and the minor ridge between the two troughs became very weak. A small, secondary Low broke off from the Phoenix Low and moved to a position in line with the deepening trough in the middle of the United States. The cyclonic vorticity with this deepening trough was in a good position for deepening of the secondary Low, and

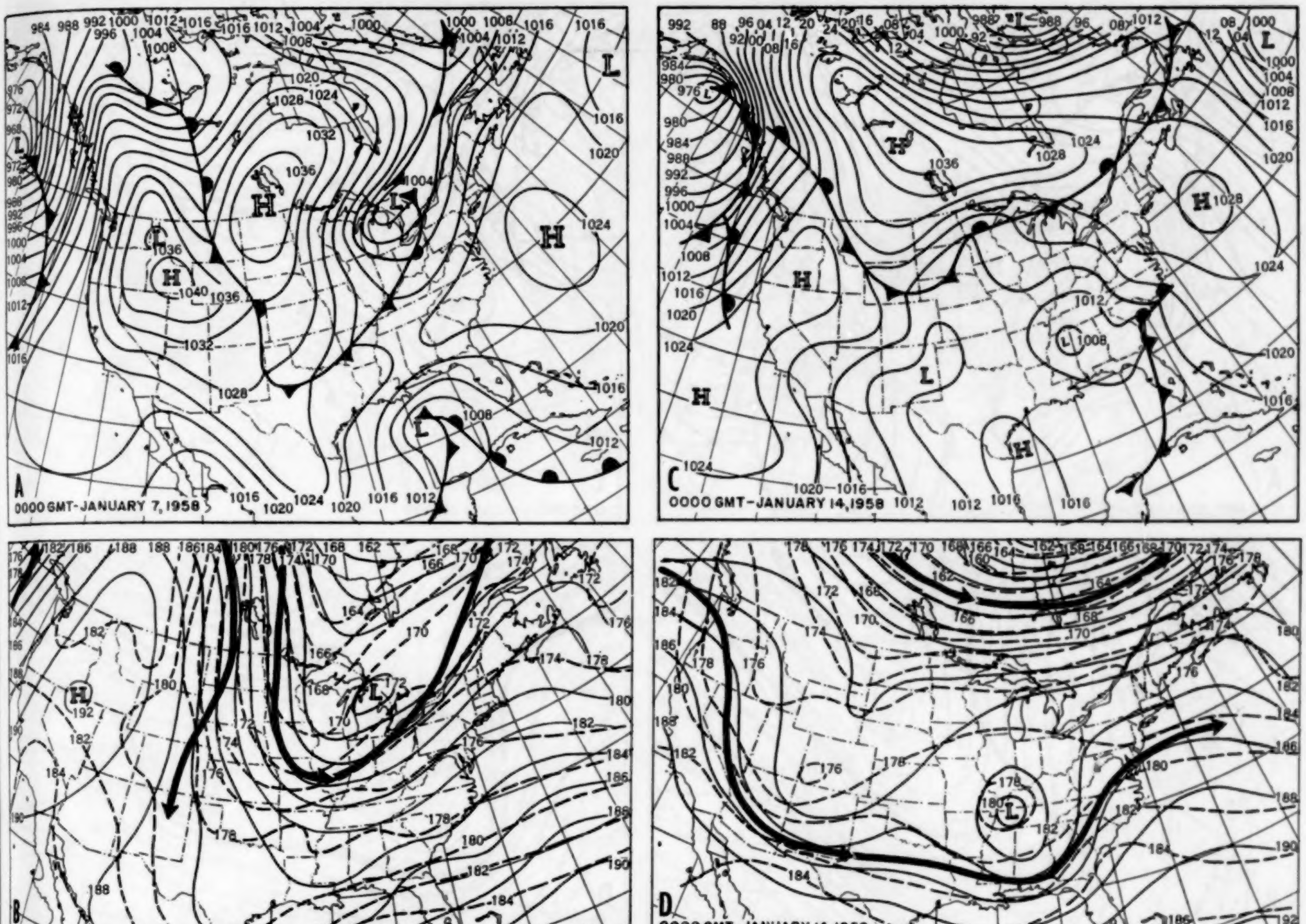

FIGURE 2.—Sea level synoptic charts (A) and (C) for 0000 GMT January 7 and 14, 1958, respectively, and the corresponding 500-mb. charts (B) and (D), with 1000-500-mb, thickness lines (dashed) and 300-mb. jets.

this is an important feature in the difference in development of these two storms.

With Storm B the western ridge became less important and the cyclonic vorticity along the western side of the trough over the central United States was negligible. The Midland Low moved eastward with cyclonic vorticity south of its center, but an objective analysis would quickly show that there was less vorticity available for surface development with this later storm. The east coast ridge remained more pronounced, but there were no strong indications of blocking action.

The 300-mb. maximum wind axes are superimposed on the 500-mb. charts to represent the upper-level jets. The main jet with Storm A entered the United States in the far Northwest, crossed Canada and dipped into the Plains States, then crossed the central Appalachians and curved east toward Bermuda. The jet with Storm B ridged across Washington State, dipped south to near Tucson, then swung east across the Southern States and ridged along the middle Atlantic coast.

3. COMPARISON OF THE STORMS

FIRST DAY FEATURES

At 0000 GMT January 7 (fig. 2A), Storm A was centered about 120 miles south of Burrwood, La.; however, there were indications that a new Low was forming southwest of Tampa, Fla. The weak to moderate wave in the Gulf of Mexico associated with this Low produced overrunning rain as far north as Atlanta, Ga. During this time, at 500 mb. (fig. 2B), a small Low was centered near Sault Sainte Marie, with a trough to the southwest over the Midwest, Mississippi, Louisiana, and coastal Texas. A weak cyclonic vorticity maximum associated with Storm A was centered over Mississippi. A second cyclonic vorticity maximum, the remnant of the Phoenix Low, was just south of Bryan, Tex. A strong southwesterly flow ahead of the trough extended from the Gulf along the Atlantic coast to the Maritime Provinces. A strong, deep northerly flow that extended from the Northwest Territories to Texas was advecting cold air down into the trough and developing favorable conditions for surface deepening.

Other surface features with Storm A were: a very slowly diminishing high pressure cell east of Cape Hatteras, that had moved east from the mid-Atlantic coast; a sharp trough, with an associated cold front, from southern Greenland to a low center over the eastern Great Lakes, then southwest over the east-central United States; and

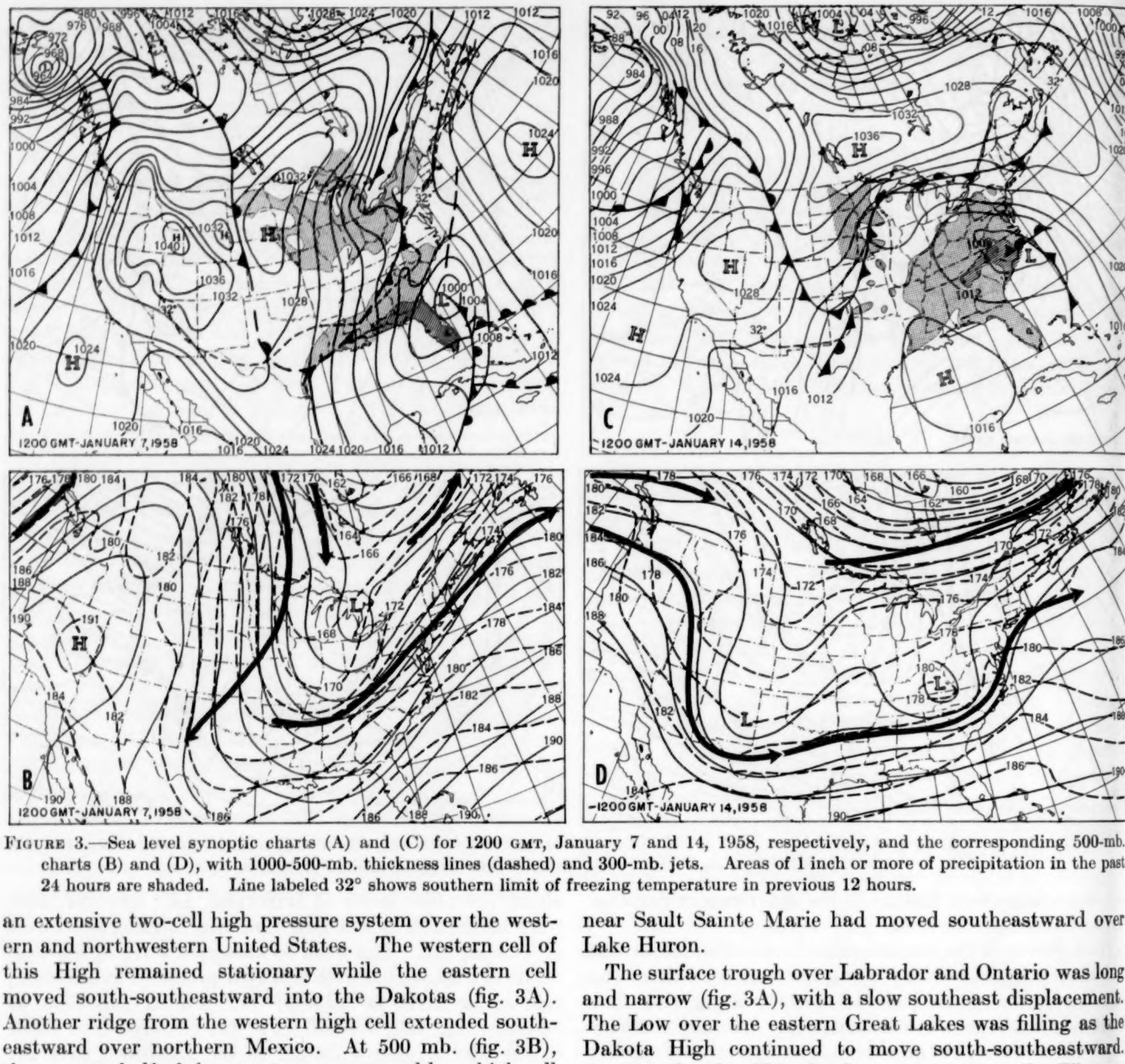

FIGURE 3.—Sea level synoptic charts (A) and (C) for 1200 GMT, January 7 and 14, 1958, respectively, and the corresponding 500-mb. charts (B) and (D), with 1000-500-mb. thickness lines (dashed) and 300-mb. jets. Areas of 1 inch or more of precipitation in the past 24 hours are shaded. Line labeled 32° shows southern limit of freezing temperature in previous 12 hours.

an extensive two-cell high pressure system over the western and northwestern United States. The western cell of this High remained stationary while the eastern cell moved south-southeastward into the Dakotas (fig. 3A). Another ridge from the western high cell extended southeastward over northern Mexico. At 500 mb. (fig. 3B), the western half of the country was covered by a high cell and ridge of great north-south amplitude with a strong southerly flow along the Pacific coast. The main jet swung up over southern Alaska, looped over far northwestern Canada and plunged south through the Plains States, crossing the central United States at about latitude 35° N., before swinging northeastward along the Atlantic coast.

By 1200 GMT January 7 (fig. 3A), a new Low had formed east of Jacksonville, Fla., with only a faint remnant of the parent Low remaining west of Tampa. The new Low reached normal intensity by this time, but its associated fronts over the ocean areas were still weak. The 500-mb. chart for 1200 GMT (fig. 3B), showed that the weak cyclonic vorticity center with Storm A had moved from Mississippi to southern Georgia, with the second cyclonic vorticity maximum over southern Louisiana. The Low near Sault Sainte Marie had moved southeastward over Lake Huron.

The surface trough over Labrador and Ontario was long and narrow (fig. 3A), with a slow southeast displacement. The Low over the eastern Great Lakes was filling as the Dakota High continued to move south-southeastward. At 500 mb. (fig. 3B), the flow pattern over the Western States remained the same, although the High over southern Utah had started weakening and the ridge over the Northwest Territories had begun to lose some of its amplitude.

At 0000 GMT January 14 (fig. 2C), a similar redevelopment took place with Storm B. While the parent Low was decreasing in intensity in northern Alabama, a new low center was forming in the vicinity of Charleston, S. C. This new Low reached normal intensity during the day (fig. 3C), while the original Low moved up the west side of the Appalachians and filled. At 500 mb. (fig. 2D), the low center with Storm B was just west of Nashville, Tenn., with a ridge to the east over the Atlantic coast. A flat trough covered the southern United States from Arizona east to the Atlantic, with a cyclonic vorticity maximum

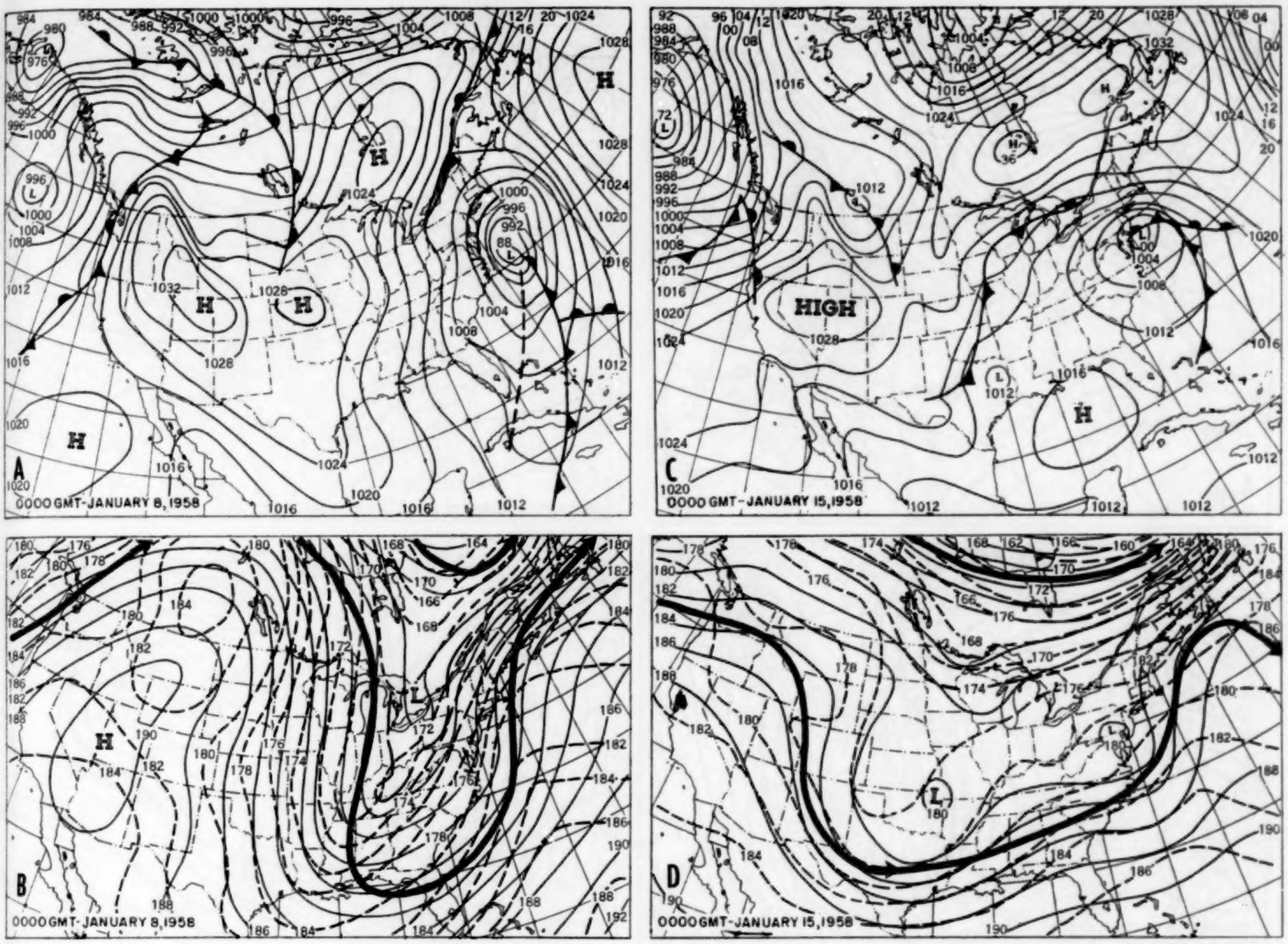

FIGURE 4.—Sea level synoptic charts (A) and (C) for 0000 GMT, January 8 and 15, 1958, respectively, and the corresponding 500-mb. charts (B) and (D) with 1000-500-mb. thickness lines (dashed) and 300-mb. jets.

over northern New Mexico, reflected on the surface by a weak Low over the Texas Panhandle.

On the surface a weak high cell was off the Atlantic coast northeast of Cape Hatteras (fig. 2C). An occlusion extended from southern Greenland to Newfoundland, then west as a stationary front roughly along the border of Canada and the United States. A high pressure cell dominated the Western States, while a broad, inverted trough covered the central United States. A High was over central Canada with a ridge to the east and southeast across New England. A small High was over the extreme western Gulf of Mexico. At 500 mb. (fig. 3D), the two centers of cyclonic vorticity moved eastward, with the New Mexico center moving more rapidly. A flat ridge, with a westerly flow, moved in over the Western States. The jet associated with this system extended from Utah south across Arizona and north-central Mexico, then east across the Southern States, with a slight northeast tilt off the Atlantic coast.

During the day (fig. 3C), the surface Low moved slowly northeastward, reaching southern Delaware by the end of the day. At the same time, the high pressure area over southeastern Canada shifted east and increased slightly. The weak trough to the southwest of Storm B persisted, with a small Low moving across Texas.

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Surface and 500-mb. features with Storms A and B were relatively similar in the eastern part of the country, with the greatest dissimilarities over the Western States (figs. 2B, 3B). The synoptic charts were similar to the extent that there was a high cell northeast of Hatteras and another extensive High over the western United States with a general trough from Greenland to the Gulf of Mexico. One of the conspicuous differences in Storms A and B was in this east coast trough. With Storm B, there was actually a trough to the north and an inverted trough to the south of a ridge that extended easterly over New England from a High centered in central Canada north of the Dakotas. The east-northeasterly path of this High was also a divergence from the High in a similar position in association with Storm A. Although there was a small high cell in the extreme western Gulf of Mexico compared to the ridge over northern Mexico with Storm A, there was a decided difference in that there was a secondary trough between this High and the principal High over the Western States.

Once they redeveloped, the surface centers of Storms A and B followed similar paths, although at different speeds, and caused similar precipitation patterns along the coast. Behind Storm A at 500 mb. the very strong north-south

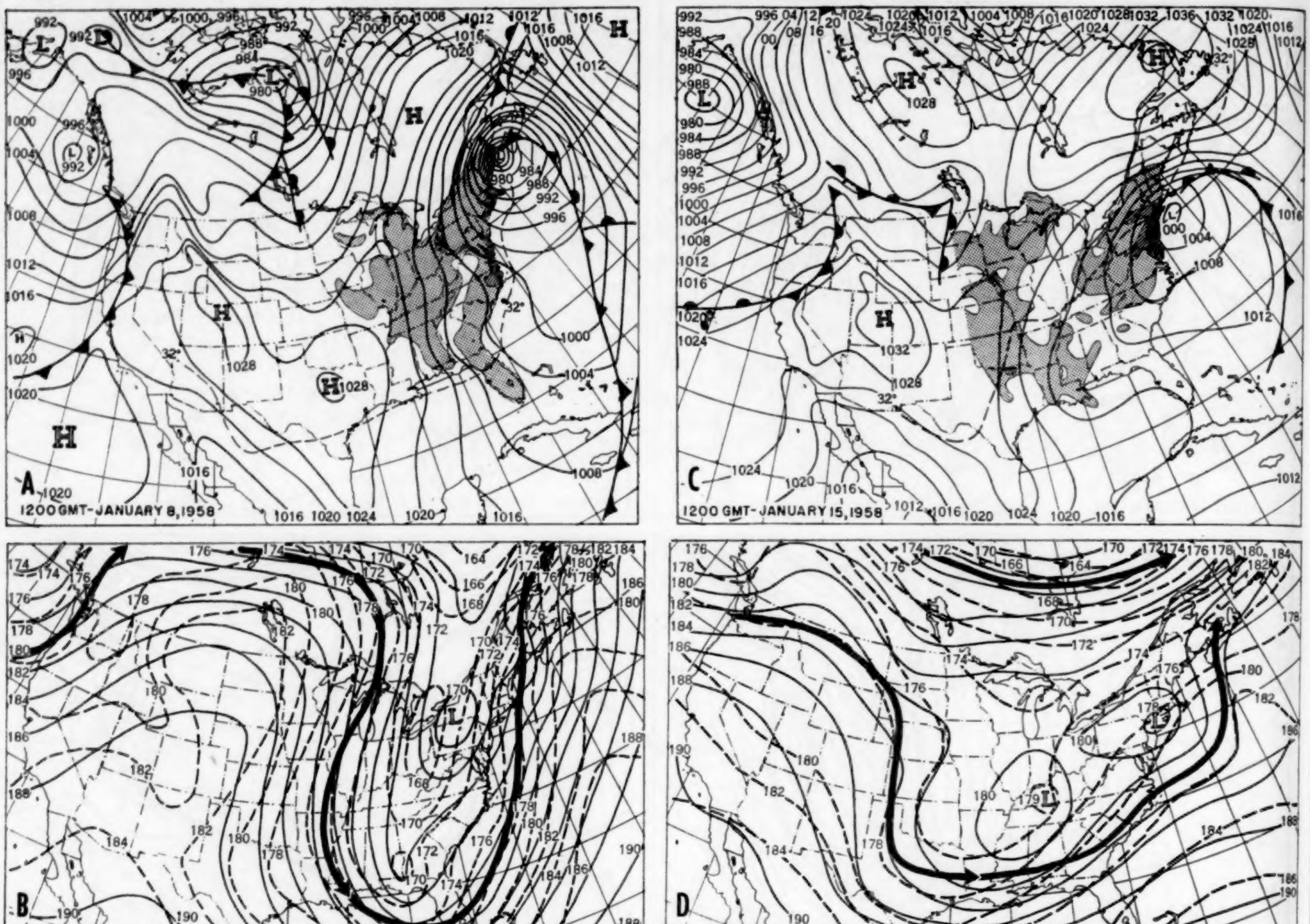

FIGURE 5.—Sea level synoptic charts (A) and (C) for 1200 GMT, January 8 and 15, 1958, respectively, and the corresponding 500-mb. charts (B) and (D) with 1000-500-mb. thickness lines (dashed) and 300-mb. jets. Areas of 1 inch or more of precipitation in the last 24 hours are shaded. Line labeled 32° shows southern limit of freezing temperature in previous 12 hours.

ridge caused a very strong northerly flow down the middle of the country. With Storm B, flow behind its low center was westerly, because of the second vorticity center over New Mexico. Storm A's cyclonic vorticity maximum was absorbed into the larger trough and was carried around it rapidly. Although Storm B's center was more definite it was imbedded in a less rapid flow and moved along its path at a more leisurely rate.

FIRST DAY WEATHER

On January 7, Storm A moved out of the Gulf, redeveloped near Jacksonville, and moved rapidly east of Cape Hatteras by 0000 GMT January 8. Rain associated with the center covered the extreme southeastern States, with amounts generally ½ to 1 inch. During the day, rain spread rapidly north and was over Nantucket and eastern Massachusetts after 1800 GMT. Snow over Virginia and inland North Carolina moved north to extend from Norfolk to the Canadian border by 1500 GMT. As rain changed to snow, sleet or freezing rain was reported in North Carolina and southern Virginia and in Providence, R. I. Snow amounts ranged from 1 to 4 inches, with heavier amounts in the immediate Atlantic coastal areas. Portland, Maine, and Hartford, Conn., reported 8 inches; Middletown, Conn., 9 inches; and Bridgeport, Conn., and Worcester, Mass., 10 inches. As the storm intensified off Cape Hatteras, winds reached gale strength over a wide area, with tides somewhat higher than usual causing minor flooding in the central Atlantic coastal area.

On January 14, the low center with Storm B was over northern Alabama. During the day it moved along the west side of the Appalachians and filled. Storm B redeveloped near Charleston, S. C., and slowly moved northward reaching southern Delaware by 0000 GMT, January 15. Precipitation with Storm B was predominantly rain, although freezing rain and sleet over Virginia early on January 14 moved up over West Virginia, Pennsylvania, and New York State during the day. Some freezing rain and sleet occurred also in southern New England and northern New Jersey. Rain spread north gradually, over the coastal areas and also along the western slopes of the Appalachians. By 0000 GMT January 15, the rain extended from eastern Kentucky to

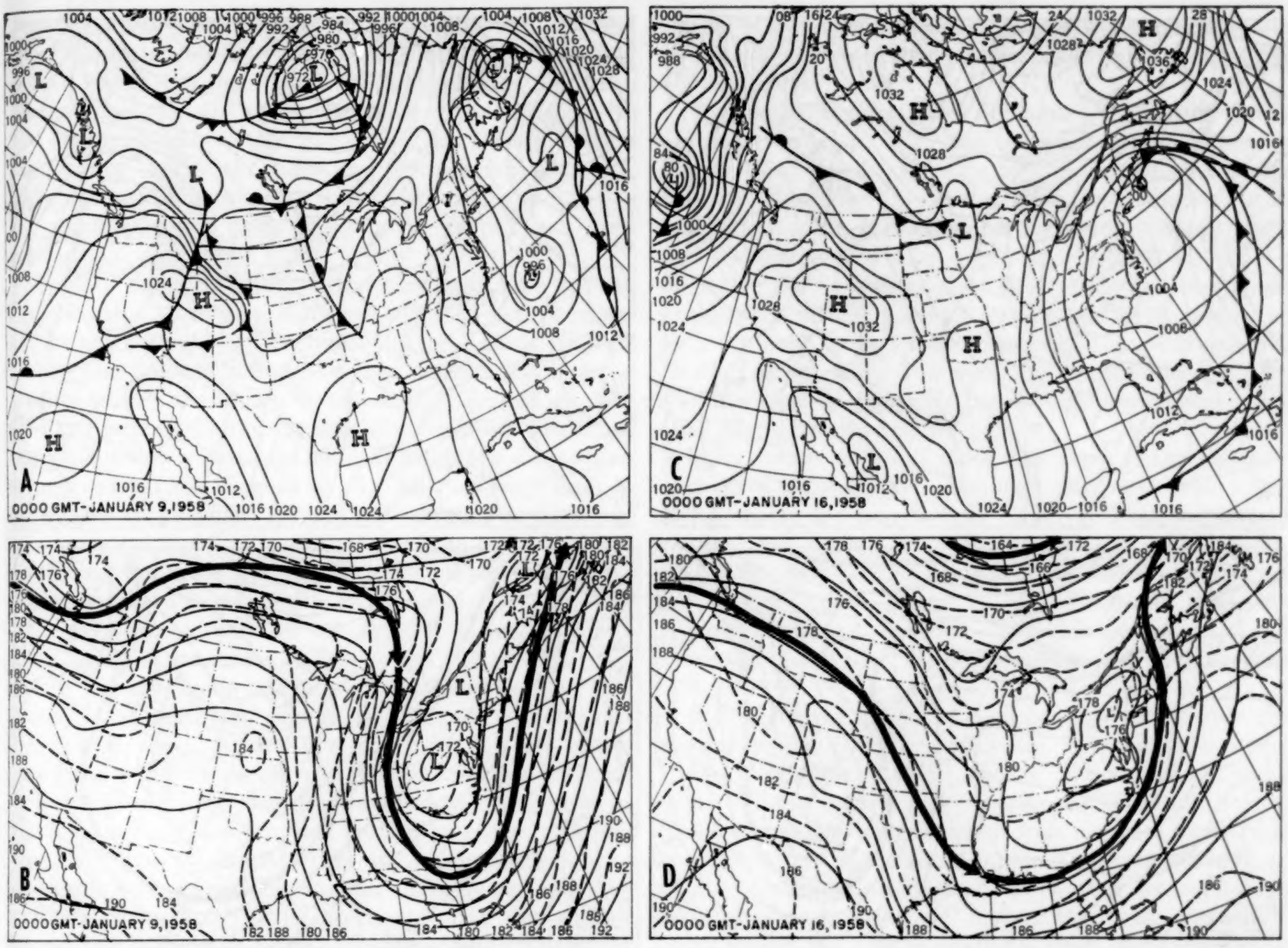

FIGURE 6.—Sea level synoptic charts (A) and (C) for 0000 GMT, January 9 and 16, 1958, respectively, and the corresponding 500-mb. charts (B) and (D) with 1000-500-mb. thickness lines (dashed) and 300-mb. jets.

Massachusetts and New York State. Storm B moved slowly, winds were not unusually strong, and precipitation amounts were not extremely large, generally ranging from ½ to 1 inch. Block Island reported 1.52 inches, Cape Hatteras 1.33 inches, and Norfolk, Va., 1.12 inches. Late in the day snow began to fall over south-central New England, with amounts ranging from 1.0 inch at Concord, N. H., and Portland, Maine, to 2.4 inches at Albany, N. Y., and 3.0 inches at Worcester, Mass.

At the end of the first day of their respective redevelopments, Storms A and B were both situated near the middle Atlantic coast—Storm A 120 miles east-northeast of Cape Hatteras and Storm B over southern Delaware. Their paths were drawing closer together—although Storm A had redeveloped near Jacksonville and Storm B near Charleston, S. C. Storm A had a more definite character and more intense circulation throughout these early stages but Storm B gradually took on a definite circulation, though not reaching the same intensity as Storm A. Precipitation patterns were similar in area. Storm A produced larger amounts of rain and snow in the north in its early stages because of its rapid movement, intensification, and overrunning. Both Storms A and B caused heavy rain and snow over southern New England, although Storm A's actual amounts were larger than Storm B's. Storm A caused heavy snow along the mid-Atlantic coast, which Storm B did not.

SECOND DAY FEATURES

On January 8, at 0000 GMT (fig. 4A), Storm A was 120 miles east-northeast of Cape Hatteras. Its central pressure was 980 mb., reflecting a 12-hour deepening of 16 mb. This central pressure, lower than two standard deviations from normal, classed the Low as very intense. The deep cold trough associated with the Low at 500 mb. (fig. 4B), was oriented northeast-southwest from Lake Erie to the western Gulf, with a center north of Toronto, Ontario. There was one center of cyclonic vorticity over northern West Virginia and another, more pronounced, over northern Alabama.

During the day, Storm A traveled northeastward at 40 knots and continued to deepen rapidly. At 0650 GMT, as it passed near Nantucket, surface pressure there dropped to 960 mb., a new record for the lowest January surface pressure. This was more than 5 standard deviations below normal for this latitude. James [2] classed Lows with central pressure more than 2 standard deviations below normal as very intense. It is interesting to

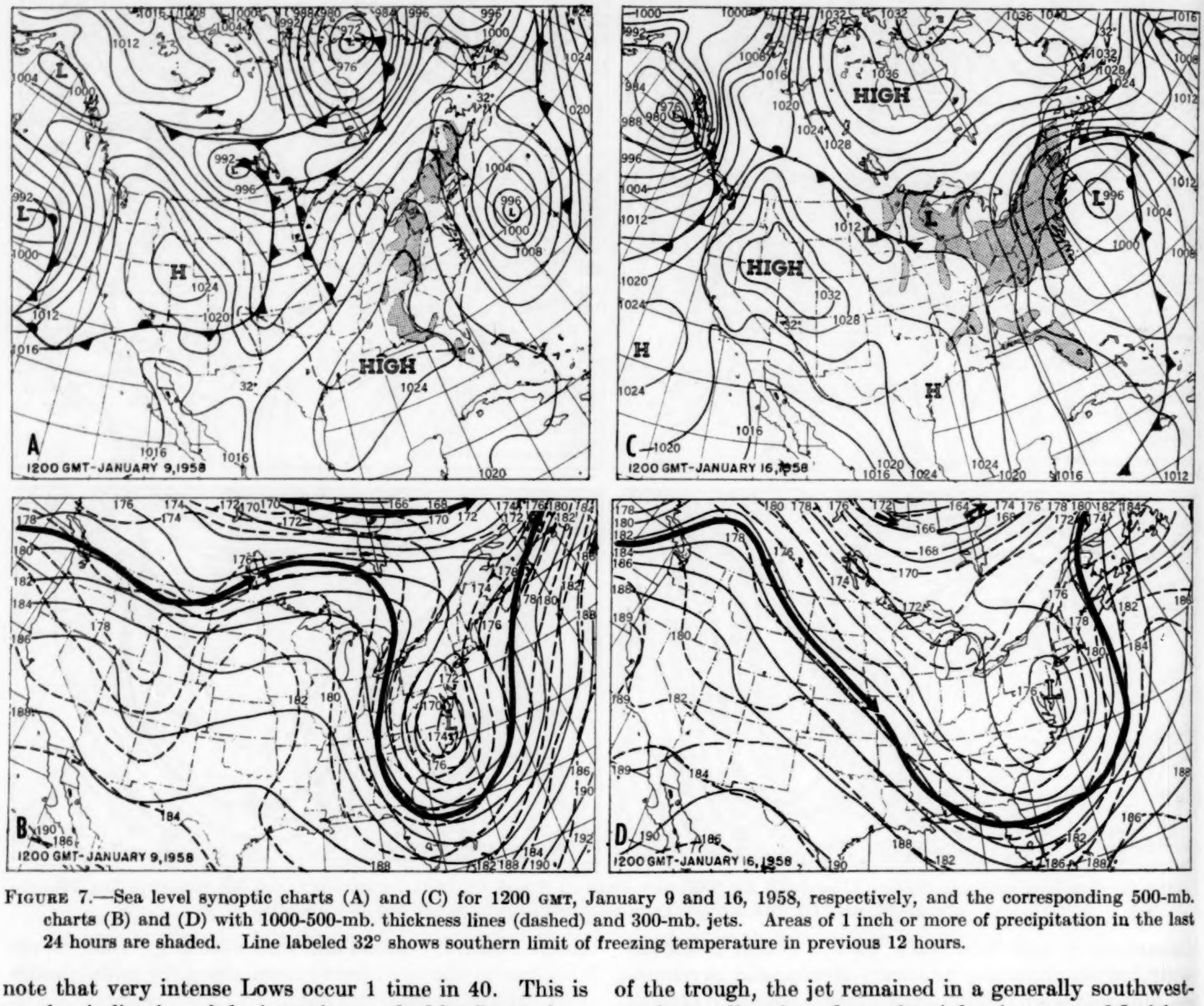

FIGURE 7.—Sea level synoptic charts (A) and (C) for 1200 GMT, January 9 and 16, 1958, respectively, and the corresponding 500-mb. charts (B) and (D) with 1000-500-mb. thickness lines (dashed) and 300-mb. jets. Areas of 1 inch or more of precipitation in the last 24 hours are shaded. Line labeled 32° shows southern limit of freezing temperature in previous 12 hours.

note that very intense Lows occur 1 time in 40. This is another indication of the intensity reached by Storm A.

At 500 mb., the cyclonic vorticity center in West Virginia moved northeastward to northern Maine (fig. 5B), and the second dropped sharply southward from Alabama to the vicinity of Albany, Ga. (fig. 5B), then moved northeastward and at day's end was off the coast of South Carolina. A residual Low, not noticeably reflected in the flat gradient at the surface, was centered near Syracuse, N. Y., and later filled (fig. 6B) as the Low formed over western Maryland.

In the 24-hour period preceding the greatest intensity of Storm A, the cold trough at 500 mb. underwent dramatic deepening (figs. 3B and 5B). The main jet dipped sharply south over the Mississippi River and across the northern Gulf. Cold air continued to be advected southward rapidly. Over most of the extreme southeastern States, 24-hour 500-mb. height falls ranged from 600 to 800 feet, with falls in excess of 800 feet at Jacksonville and Eglin AFB, Fla. Correspondingly, temperatures at the 500-mb. level dropped 10 to 12 C.° in 24 hours, with a 14° drop at Patrick AFB, Fla., and a 15° drop at Jacksonville. Ahead of the trough, the jet remained in a generally southwest-northeast direction along the Atlantic coast, Maritime Provinces, and Newfoundland, and carried warm air well to the northeast.

As Storm A moved away from the Cape Cod area, it began filling (fig. 5A), and at 0000 GMT January 9 (fig. 6A), had a central pressure of 986 mb. As it filled, the 500-mb. Low lost its closed circulation and became absorbed into the trough, moving by day's end, to the Gulf of Saint Lawrence. As Storm A moved northeastward, a surface trough remained to the southwest behind it and a second, quasi-stationary, Low formed 250 miles southeast of Cape Hatteras. During this time, the low center at 500 mb. traveled from eastern Tennessee to western Maryland, deepening steadily until it became the dominant 500-mb. Low over the eastern United States.

At the surface, while Storm A traveled through the Atlantic coastal waters the Atlantic high cell south of Gander, Newfoundland, moved slowly eastward with a 4-mb. increase in its central pressure. The weak ridge associated with this High at 500 mb. moved eastward also, and weakened slightly. The large surface High persisted

over the Northwestern States while the ridge of high pressure associated with it aloft weakened, flattened, and moved from Utah (fig. 4B) to the Plains States (fig. 5B). The main jet (fig. 4B) at the beginning of the day, crossed the extreme northwestern portion of the United States, then swung well to the north over central Canada before dipping south. As the ridge weakened during the day, the jet remained over Canada but took on an east-west orientation.

A sharp ridge built up to the northeast over central Ontario, near Moosonee. This was reflected at the surface by the quick southward movement of the Kansas High to Texas, and the persistence of a long, narrow ridge across south-central Canada east to Goose Bay, Labrador. Behind the surface High in the west, the occlusion weakened and drifted slowly eastward while at 500 mb. a short-wave trough entered the Northwest (fig. 6B). A very intense low center over Hudson Bay moved eastward as its upper trough moved rapidly across the top of the 500-mb. ridge. A surface warm front associated with the Hudson Bay Low moved eastward across North Dakota.

On January 15, at 0000 GMT (fig. 4C), Storm B was about 20 miles southeast of Dover, Del. A central pressure of 996 mb. made it a normal to intense Low for this latitude, and there was very little change in its intensity through the day. At 500 mb. (fig. 4D), the low center associated with Storm B was about 25 miles west of Philadelphia. During the day it moved slowly northward while the surface Low traveled northeastward at 15 knots to reach Boston by the end of the day. Flow ahead of the storm was easterly and increased in intensity as the center moved northward. The jet associated with this system crossed Texas and the South, then swung northeastward along the Atlantic coastline.

A weak trough remained to the south of Storm B, between the Atlantic coast and Bermuda. Another trough, though less pronounced, extended southwest to a small low center over eastern Texas (fig. 4C). These troughs caused a northeasterly return circulation over the northeastern United States (fig. 5C). The small eastern Texas Low was in conjunction with a 500-mb. Low centered near Joplin, Mo. (fig. 4D). During the day, as the surface Low traveled east to become absorbed into the larger Atlantic coast trough, the upper Low moved very rapidly eastward (fig. 5D), to join with the Low associated with Storm B. This double Low system at the 500-mb. level caused a wider trough aloft, with a flatter, more westerly flow across the Southern States. The 500-mb. ridge over the Northwest (fig. 4D), was also relatively flat, so the circulation pattern did not produce any marked warm or cold advection.

The weak front across the North Central States and the Plains States dissipated during the day (fig. 5C). The surface ridge in the Atlantic east of Storm B persisted, as the Moosonee High moved eastward off the coast of Labrador and Newfoundland, with a 5-mb. increase in pressure. At 500-mb., the Atlantic ridge moved very slowly eastward, with no appreciable change in intensity. As the Moosonee High moved eastward a strong westerly circulation was evident over southeastern Canada at the 500-mb. level (fig. 5D).

The surface high cell over northern Utah drifted slowly southeastward (fig. 5C) as an occlusion moved in behind it and frontolyzed. A ridge to the east of this center built up during the day over an area from Texas north to Nebraska, and a lee trough formed over western Texas and eastern New Mexico. At 500 mb. (figs. 4D, 5D), a moderate ridge persisted over the Northwest throughout January 15, although a very weak short wave, associated with the weak trough and warm front over Montana and North Dakota, moved in over the Northwestern States. The main jet was westerly across Washington and Montana (fig. 4D).

Both storms traveled along the same paths on their respective second days, although Storm A was much more intense and moved more rapidly. Both storms had residual low centers form in the troughs behind them. Although there were quite a few similarities in the surface patterns of Storms A and B, at the 500-mb. level, the greatest similarity was in the positions and tracks of the Lows associated with the main surface centers of the two storms.

At the surface, ridges of high pressure set in over the central United States in both cases, although with Storm B the ridging process was slowed down while the small eastern Texas Low traveled across the Southern States. The 500-mb. flow over the ridge with Storm A was strong and northerly; with Storm B, the flow was westerly of moderate strength, and gradually shifted around to become more northwesterly. Both of these ridges extended into Canada, but with Storm B the ridge had an eastward elongation. The High centered off the Labrador coast with Storm B was more intense and became a formidable block.

Surface pressure was generally high over the Northwest with associated ridges aloft over the Western States. With Storm A the ridge was a large-amplitude ridge that flattened during the day, while Storm B had a flatter ridge over the western half of the country throughout the day.

The great amplitude of the 500-mb. ridge and trough with Storm A produced very strong advection of cold air behind the trough and warm air ahead, contributing importantly to its very dramatic deepening. With Storm B, the amplitude of the ridge and trough at 500 mb. was less accentuated and no pronounced warm or cold advection was associated with the system.

SECOND DAY WEATHER

On January 8, as Storm A swept northward, rain was general along the coast, with snow over North Carolina and Virginia. During the day, snow spread gradually over northern Virginia, eastern Maryland, Delaware, and New Jersey reaching later in the day over Pennsylvania, New York State, and New England. Sleet and freezing rain occurred spottily in a line roughly through Connect-

icut, northern Pennsylvania, central New York State, and even into Ohio. Snow occurred over the Appalachians, Kentucky, Tennessee, and well into central Alabama. Snow amounts along the immediate coastal area were heavy, with amounts over 12 inches in eastern Maryland, and 8 to 15 inches along the New Jersey coast. Over New England, snow amounts had accumulated 4 to 12 inches, and along a 30- to 50-mile-wide belt from north-central Maine to the Connecticut coast accumulations totaled 12 to 20 inches. Portland, Maine, and Bridgeport and Hartford, Conn., reported 11.0 inches of snow; Middletown, Conn., 12.0 inches; Worcester, Mass., 13.0 inches; and Concord, N. H., 14.5 inches. Snow was reported at several locations in Florida and Georgia, although it melted as it fell or left just a small amount on the ground. Heavy amounts of rain fell over Cape Cod, Massachusetts' south shore, and the eastern Long Island-Nantucket area. Nantucket reported 1.56 inches of rain with the storm, Block Island 2.31 inches, and the Cape Cod area amounts up to 4.50 inches. Storm A was accompanied by violent winds of hurricane force near its center as it passed over the Nantucket and Cape Cod areas. Considerable damage was caused by the winds and resulting higher than normal tides.

On January 15, as Storm B moved slowly north, it was preceded by general rains covering southern New England, New York State, Pennsylvania, and from West Virginia east to the Atlantic coast. Rain amounts ranged from half an inch to 1 to 3 inches. Providence reported .91 inch and Boston 1.79 inches. Freezing rain and sleet occurred along Lake Erie and Lake Ontario and in scattered areas in New York and Connecticut. Snow was falling over southern Vermont and New Hampshire and the southwestern corner of Maine. During the day, snow spread gradually over New England. Snow amounts were generally 1 to 3 inches, with 5 inches reported at Portland, Maine, 6 inches at Burlington, Vt., 7 inches at Albany, N. Y., and 9 inches at Caribou, Maine. As Storm B moved over the New England coastal area its center was accompanied by winds of gale force—the fastest mile of wind at Nantucket was 52 m. p. h. from the east. As colder air flowed down behind the storm, precipitation over Ohio, Pennsylvania, and West Virginia changed to snow. At the same time, snow and some areas of freezing precipitation spread into Kentucky and Tennessee.

Although both Storms A and B had marked similarities in their paths, precipitation patterns, and precipitation amounts, there was some difference in their intensity. Storm A produced heavy amounts of snow and rain along the coastal areas and as it reached its lowest pressure produced hurricane winds near its center. Its movement was very rapid and weather associated with it was severe, in some areas causing considerable damage. Precipitation associated with it reached from central Alabama across the whole tier of Eastern States. Storm B produced large amounts of snow and rain also, but its progress was slower, and its precipitation, though widespread, not generally of such large extent or amounts as those of Storm A. Winds reached gale strength with Storm B, but did not cause damage as serious as that of Storm A. Storm B's slower movement was another cause of its large precipitation amounts.

THIRD DAY FEATURES

The phase of development on the third day represented the concluding differences in the storms. By 0000 GMT January 9 (fig. 6A) Storm A had filled to 986 mb., which classed it an intense Low, as it raced northeastward across the Atlantic Ocean. At 500 mb. (fig. 6B) the Low was north of Nova Scotia, a part of the extensive trough from Labrador southwest to the strong Low over western Maryland. During the day (fig. 7B), this strong Low center moved to the southeast. A rather straight southwesterly flow was ahead of this trough while the broad area of northerly winds to the west remained. The Nova Scotia Low was reflected in the very intense residual Low that had developed east of Cape Hatteras (fig. 6A). This Low persisted, and as it slowly filled, moved northeastward along the track Storm A had followed (fig. 7A).

The High in the Atlantic gave way slowly to the east and the ridge pushing into the western Gulf of Mexico amalgamated with the high cell that had traveled south from Canada to form a 1027-mb. High in the western Gulf (fig. 7A). This regime continued throughout the third day, with a large easterly-moving Low in central Canada and its companion wide trough lagging to the lee of the Rockies. The Western States remained under the influence of a large stationary high cell over northern Utah.

As Storm A's 500-mb. Low moved out and the residual Low took over, the marked cold air advection decreased. To the west, a 500-mb. short-wave trough associated with the surface trough over the Plains States, moved in along the westerly flow (fig. 7B). The main jet dipped down over Washington and Montana in association with this short-wave trough, then swung east across south-central Canada, south around the residual Low, and then northeast over the Atlantic and Newfoundland.

At 0000 GMT January 16 (fig. 6C) the blocking High had established itself well along southern Canada and Storm B almost came to a standstill. Minor new developments to the south of the Low moved up to the block and were absorbed into Storm B's original Low (fig. 7C). Instead of filling on the third day as Storm A did, Storm B continued to intensify (fig. 7C) reaching very intense status near the end of the period. However, Storm B never reached an intensity as great as that of Storm A.

At 500 mb. (fig. 6D) a marked sharp ridge extended along longitude 60° W., from latitude 35° N. to the Maritime Provinces. The two centers of cyclonic vorticity in the east consolidated (fig. 7D) and moved eastward. A moderate northwesterly flow persisted behind this eastern trough during the day, as a minor trough crossed the Midwest.

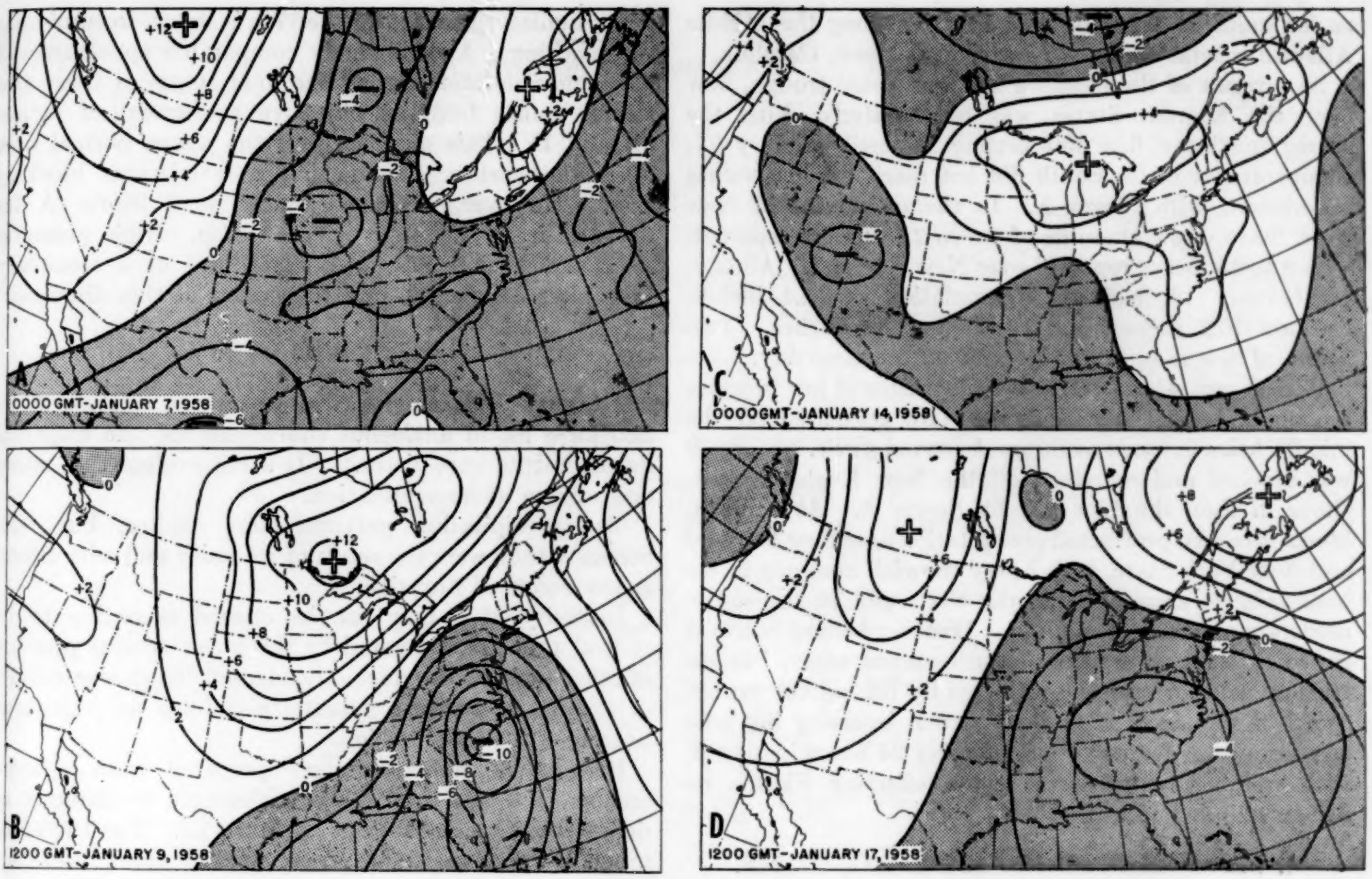

FIGURE 8.—Departure from normal of the 1000-500-mb. thickness, (A) 0000 GMT, January 7, (B) 1200 GMT, January 9, (C) 0000 GMT, January 14, and (D) 1200 GMT, January 17, 1958.

At the surface at day's end (fig. 7C) Storm B had begun to retrograde to the northwest. A northerly flow covered the Eastern States, with a stationary High over the Far West and a second High over Texas. A weak, complex low pressure area was over Minnesota and the Dakotas.

On the third day, Storm A moved away to the northeast while Storm B was blocked by the high pressure cell in the Labrador area. Troughs persisted to the south behind both storms, causing cold northerly flow over the Eastern States. In both cases stationary Highs dominated the West with high pressure areas over Texas and the western Gulf and low pressure areas over the Plains States.

THIRD DAY WEATHER

On January 9, at 0000 GMT (fig. 6A), Storm A had moved over Newfoundland. A second small low center, of normal intensity (999 mb.), was located 5° of longitude east of Cape Hatteras. The strong northerly circulation had set in over the Eastern States. Snow flurries were general over New York State, upper New England, to the lee of Lake Erie, and along the Appalachians. During the day, Storm A moved rapidly eastward and the secondary Low drifted very slowly to the east. Temperatures in the South and East which were under the influence of cold air for several days, dropped well below normal. Temperatures in north and central Florida on January 9 ranged from 18° to 29° F. and in the Everglades from 26° to 30° away from the lake and 30° to 35° near the lake. Temperatures along the east coast were above freezing, but inland, freezing temperatures and frost were reported. The cold weather caused damage to citrus fruit, especially in the north, and heavy damage to truck crops and young plants. Some minimum temperatures in Florida on January 9 were: Tampa 31° F., Lakeland 28°, Orlando 27°, Apalachicola 26°, and Tallahassee 23°.

Elsewhere in the South, temperatures ranged from near zero (5° F. at Farmerville, La.) through the teens and twenties (Greensboro, N. C., and Rome, Ga., 10° F.; Norfolk, Atlanta, and Birmingham 17°; Shreveport and Baton Rouge 23°; and New Orleans, 26°). Although this cold spell set no great number of extreme minimums, the cold air covered a broad area and persisted for 2 to 4 days, where usually such cold spells move quickly off after 1 or 2 days.

On January 16 (fig. 7C), as Storm B remained off Cape Cod, blocked by the large High over Labrador, a trough extended to the southwest along the Atlantic coast. Snow fell over Vermont, New Hampshire, western Maine, New York State, and Pennsylvania. Rain occurred in the coastal areas of New England and southeastern Maine. Snow flurries fell along the Appalachians

as far south as Asheville, and rain fell along the Middle Atlantic coastal area and over central North Carolina.

As a result of the blocking and east coast trough, flow over the Eastern States was northeasterly, with the strong northerly flow not setting in until January 17. Temperatures in the South did not reach the low values experienced with Storm A. In the mountains of New York State, snow amounts of 1.5 to 2 feet were reported, with 8 to 15 inches reported over New England. Albany, on January 16, had an accumulation of 15.4 inches, Caribou 10.2 inches, and Burlington 8.9 inches. Two inches of new snow fell at Newark on the same day.

Both Storm A and Storm B had troughs of low pressure remaining behind them. A new Low generated off the middle Atlantic coast as Storm A moved away. Storm B was blocked and remained off the New England coast, closer in than the Low left by Storm A. As a result, Storm B caused protracted precipitation over New England and New York State, with heavy snowfall amounts in the mountainous areas. Along the coast and to the south, moderate amounts of rain fell. Weather behind Storm A ended as soon as the main storm moved away, but extremely cold air rushed south rapidly behind the center. Storm B remained at or near normal intensity but cold advection was delayed approximately 24 hours behind it, with freezing temperatures again affecting Florida on January 17 and 18.

4. DEPARTURE FROM NORMAL 1000-500-MB. THICKNESS

In objective evaluations of the relative intensities of storms, the thermal field is of significance. For the consideration of this concept to complement the previous impressions from the classification of intensities according to James [2], the 1000-500-mb. thickness departures from normal were examined (fig. 8). Conditions at the time of redevelopment are shown by the 0000 GMT charts (figs. 8A and C), and during the concluding stages by the last day's 1200 GMT charts (figs. 8B and D) of each of these storms. At the time of redevelopment, both storms had reached normal intensity, but Storm A obviously had the greater potential according to the departure from normal charts. Although there was a very slight above normal thermal field ahead of Storm A, the below normal area to the rear denoted a strong thermal gradient in the redevelopment area. The above normal area ahead of Storm B was more prominent in the proximity of redevelopment and gave a clue to blocking action that became of more consequence later.

The departures from normal for January 16 and 17 being similar, the chart for the 17th is shown to emphasize the blocking. A surprisingly comparable appearance of blocking conditions during the final stages of these two storms comes from an objective assessment of figures 8B and D. This assumes that the above normal area over southeastern Canada indicates a warm blocking ridge. However, rapid movement took Storm A far seaward from the area by that time. High pressures did build behind Storm A and ahead of a secondary Low after January 9, and the period of this discussion.

5. CONCLUSION

Although we have classed these storms as Texas Lows and have found analogous characteristics, the disparate developments were distinct. It is not evident that either one was an average situation.

The precipitation patterns were similar, but with Storm A they were the result of intensity and with Storm B the result of duration.

Intensity of a Low can be classed according to its central pressure compared to the mean central pressure of other Lows at the same latitude, but the characteristics of associated pressure systems also have an important bearing on its classification.

Because normal charts are prepared from a large amount of data, they reflect differences as strongly as does a study such as the one by James [2]. The 1000-500-mb. thickness departure from normal chart does not indicate the intensity of Lows, but it gives evidence of the intensity of thermal fields. In the cases presented here, the more pronounced thermal gradient, as indicated by the greater magnitude of the departure from normal, appears with the more intense storm.

ACKNOWLEDGMENTS

The authors wish to thank the staff of the National Weather Analysis Center for preparation of the charts used and the staff of the Daily Map Unit of the Weather Bureau for drafting the figures.

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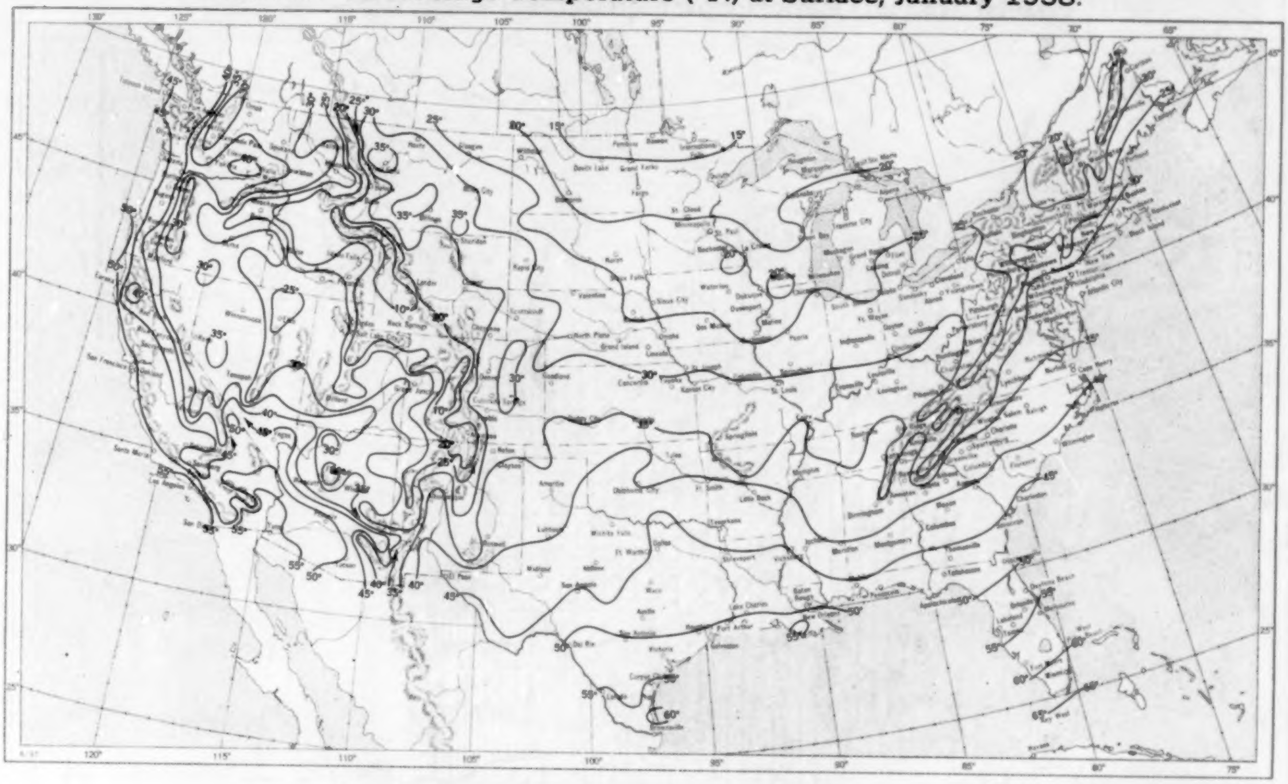
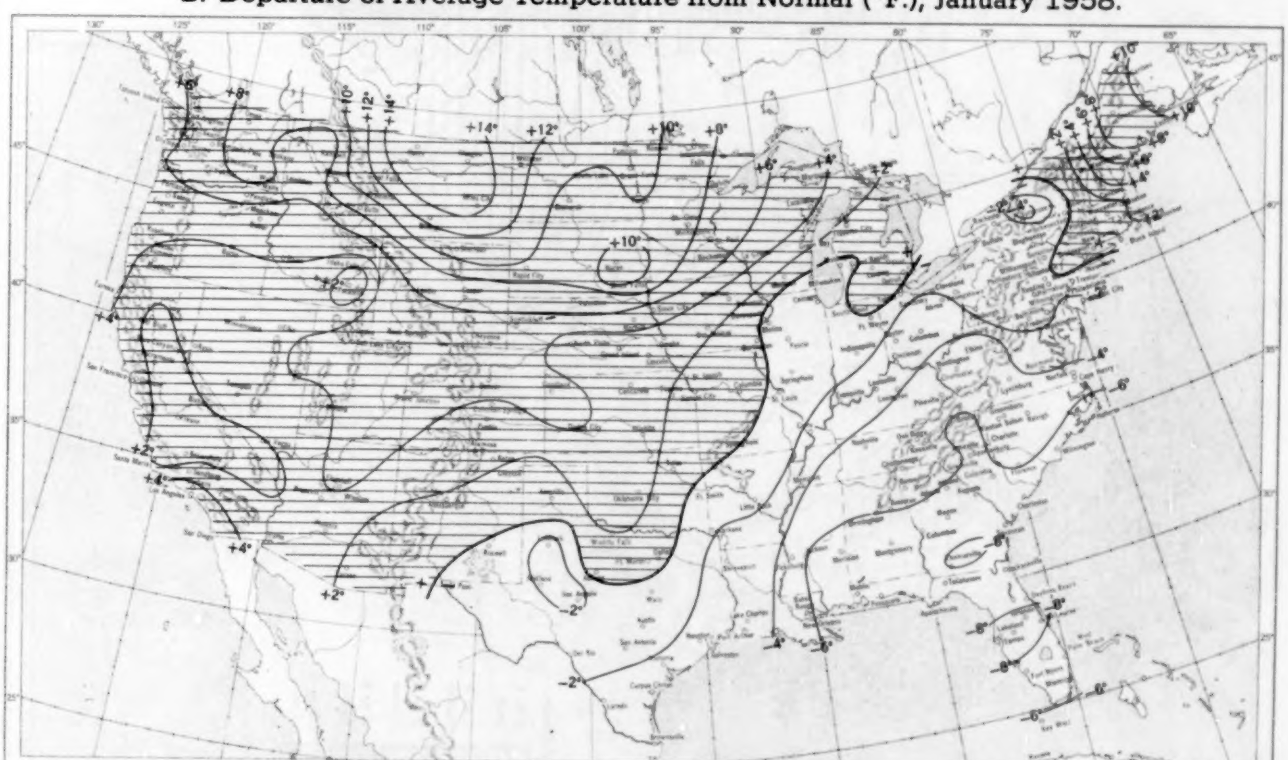
U. S. GOVERNMENT PRINTING OFFICE: 1958

Chart I. A. Average Temperature (°F.) at Surface, January 1958.

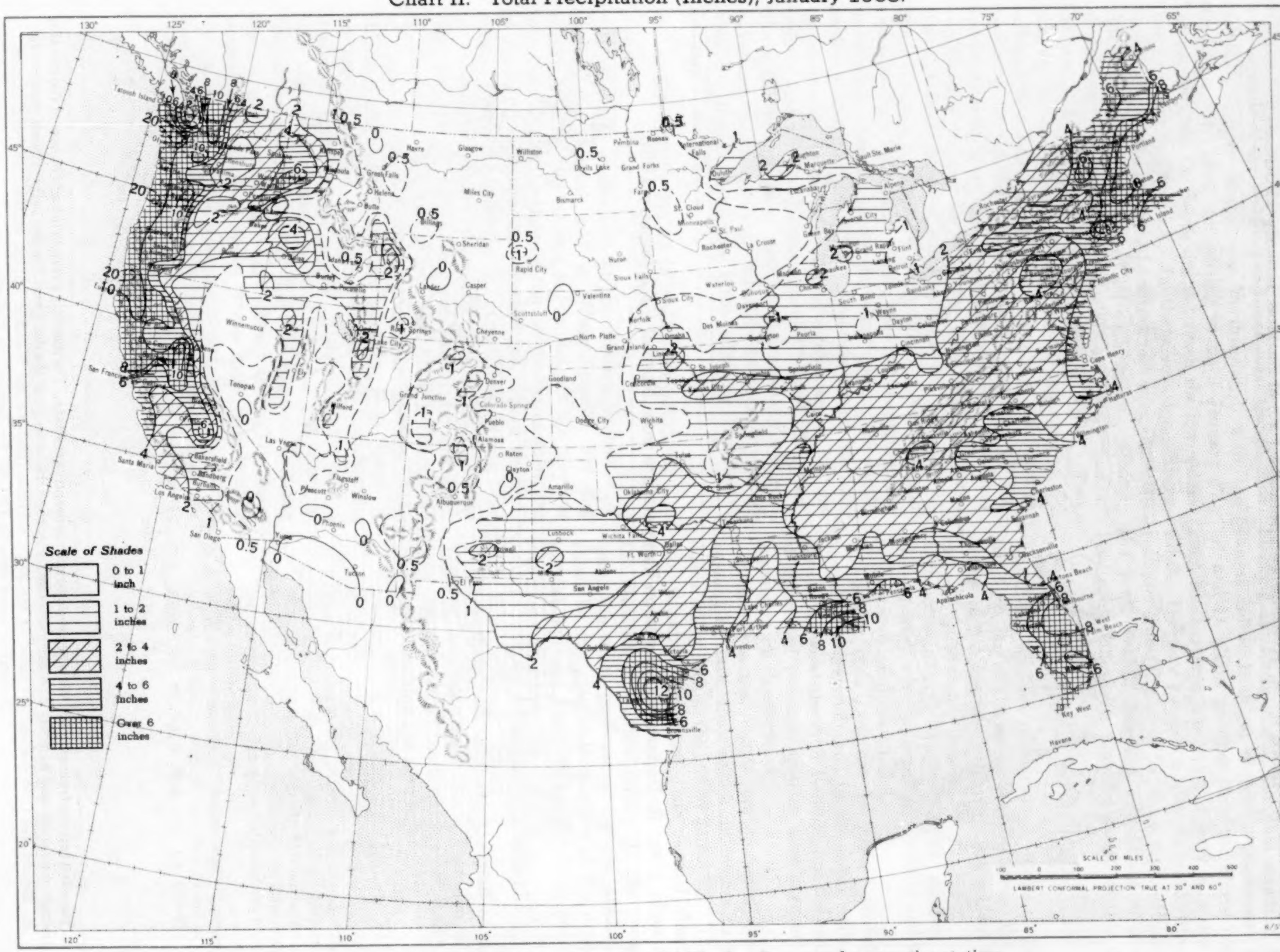


B. Departure of Average Temperature from Normal (°F.), January 1958.



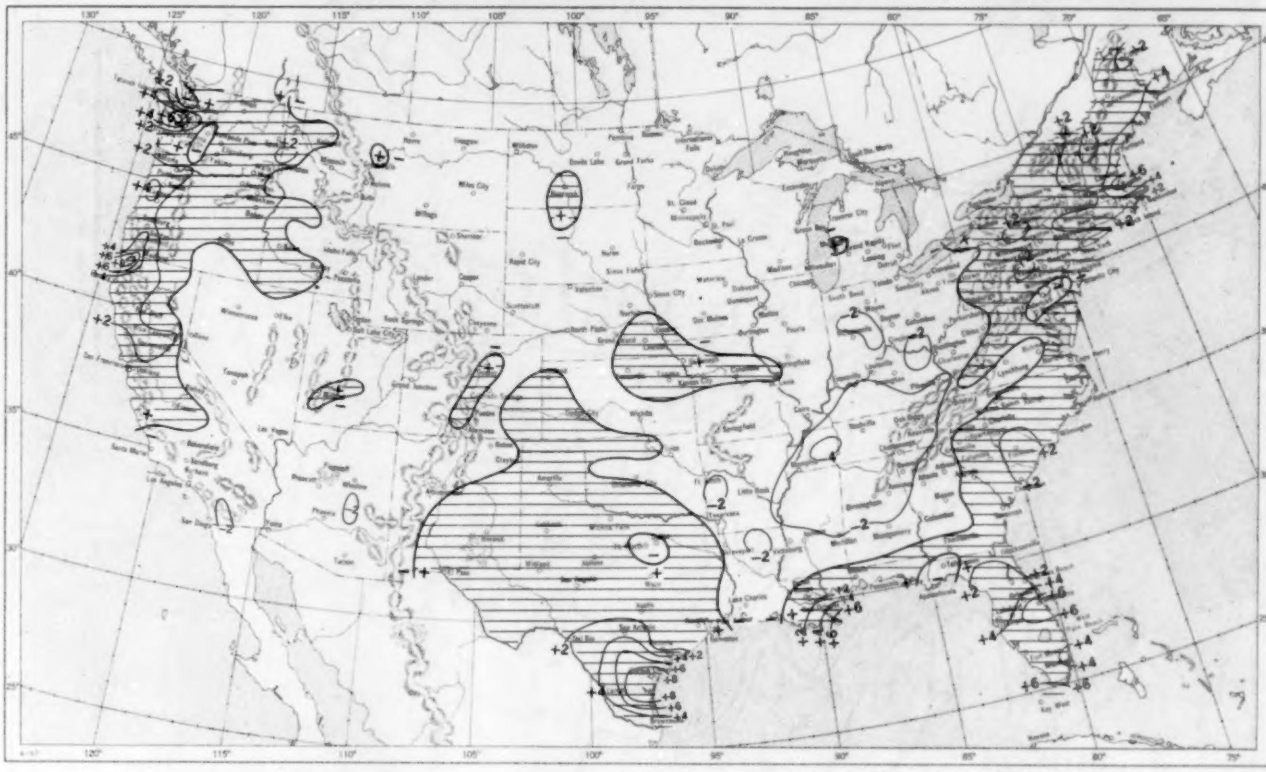
A. Based on reports from over 900 Weather Bureau and cooperative stations. The monthly average is half the sum of the monthly average maximum and monthly average minimum, which are the average of the daily maxima and daily minima, respectively.
B. Departures from normal are based on the 30-yr. normals (1921-50) for Weather Bureau stations and on means of 25 years or more (mostly 1931-55) for cooperative stations.

Chart II. Total Precipitation (Inches), January 1958.



Based on daily precipitation records at about 800 Weather Bureau and cooperative stations.

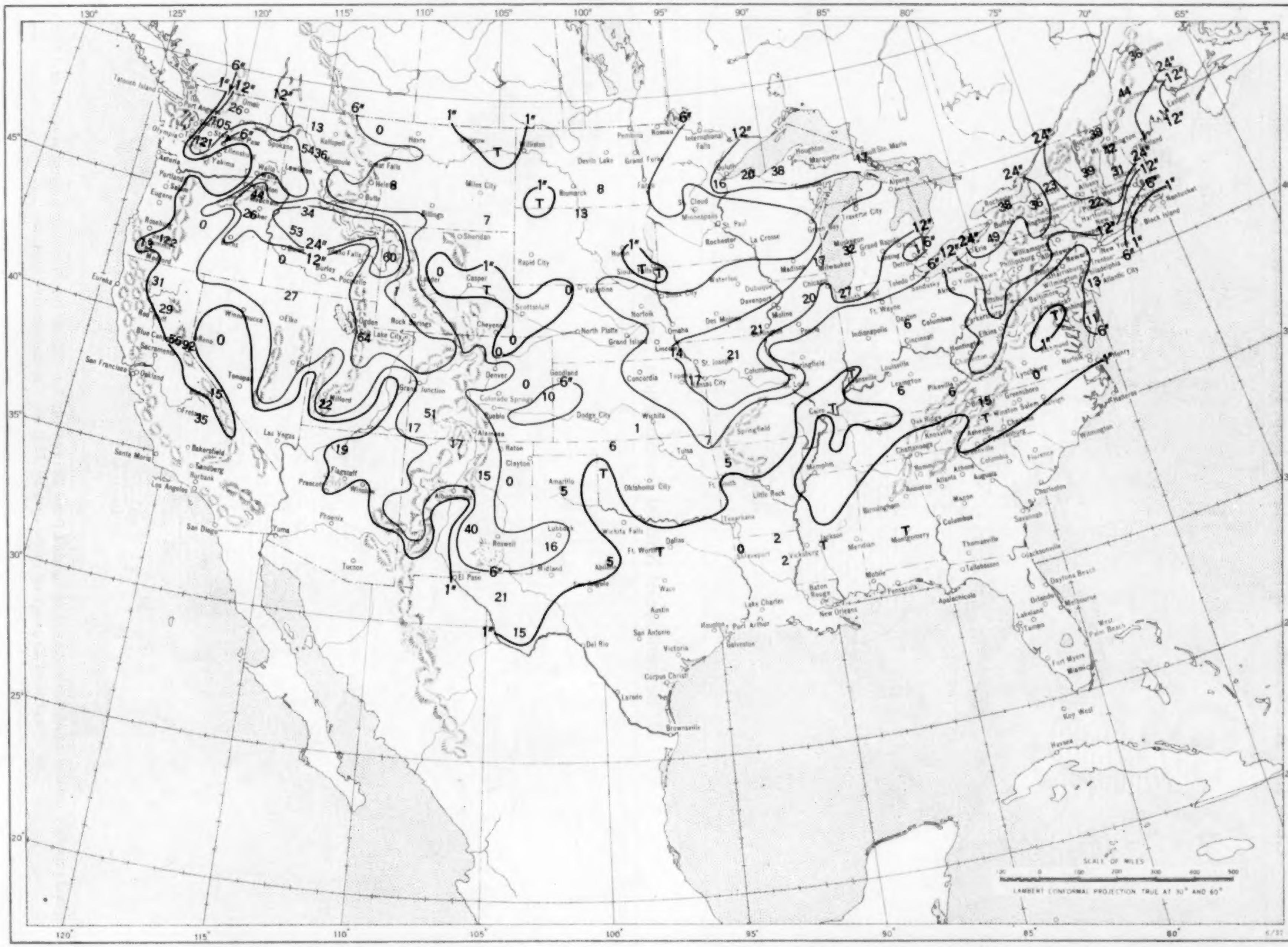
Chart III. A. Departure of Precipitation from Normal (Inches), January 1958.



B. Percentage of Normal Precipitation, January 1958.

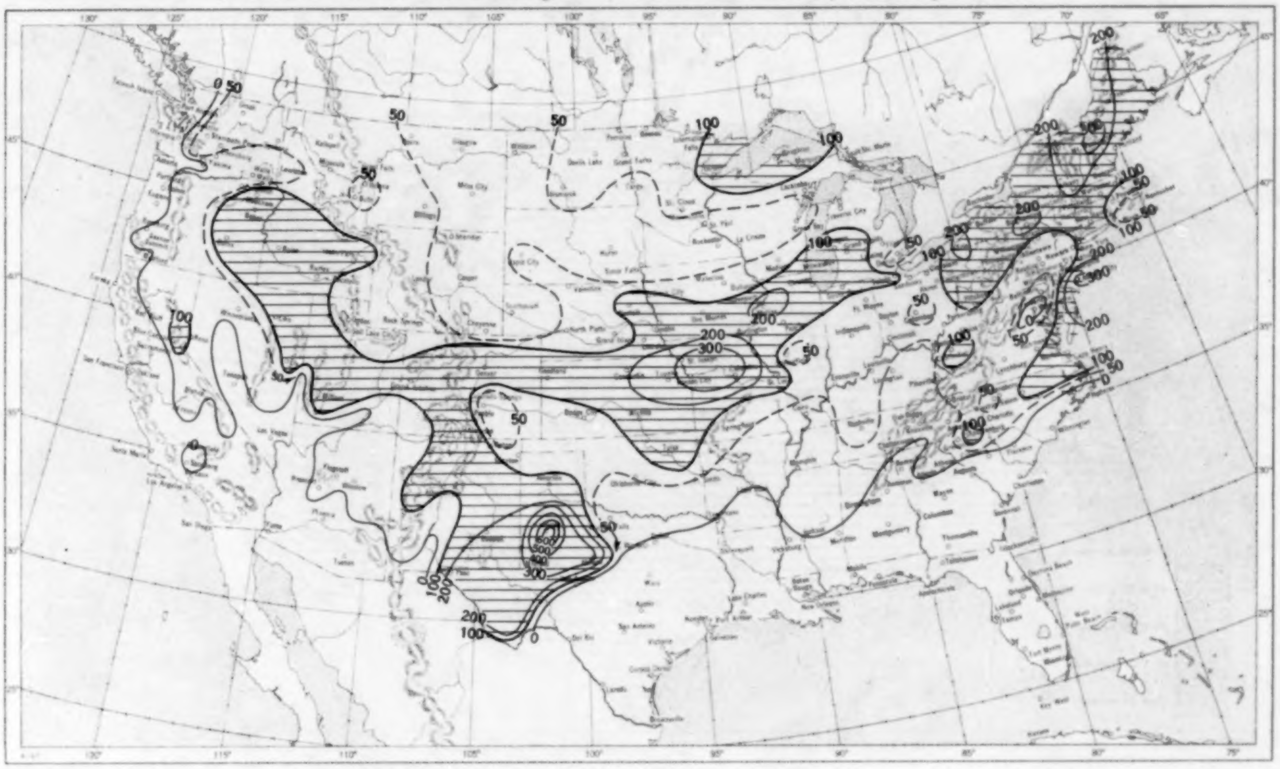
Normal monthly precipitation amounts are computed from the records for 1921-50 for Weather Bureau stations and from records of 25 years or more (mostly 1931-55) for cooperative stations.

Chart IV. Total Snowfall (Inches), January 1958.

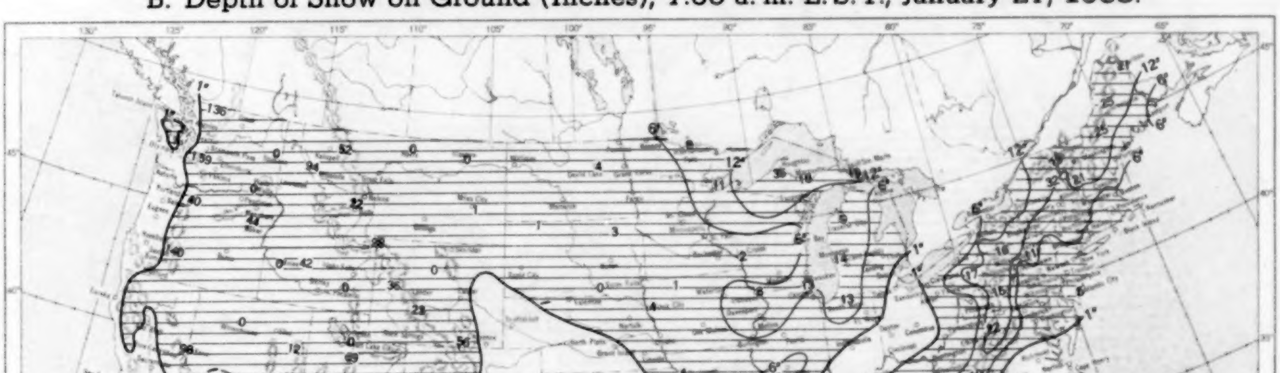


This is the total of unmelted snowfall recorded during the month at Weather Bureau and cooperative stations. This chart and Chart V are published only for the months of November through April although of course there is some snow at higher elevations, particularly in the far West, earlier and later in the year.

Chart V. A. Percentage of Normal Snowfall, January 1958.

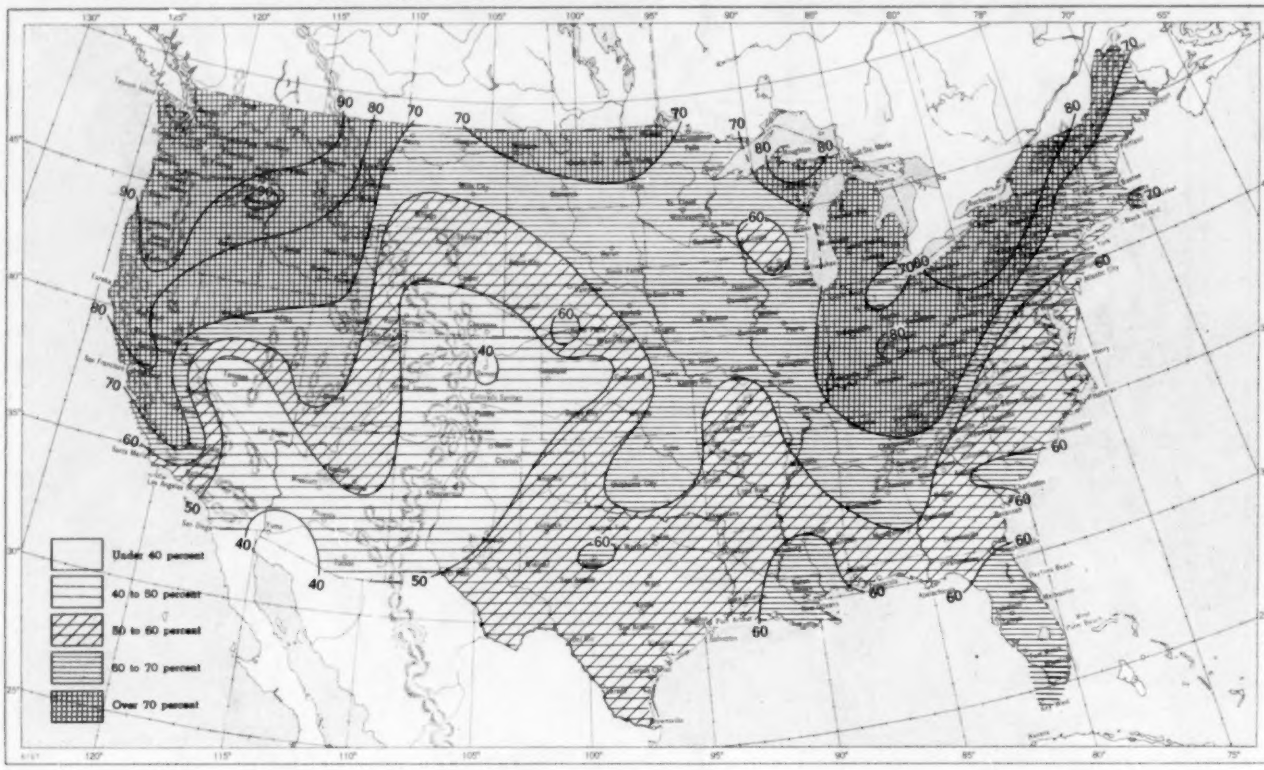


B. Depth of Snow on Ground (Inches), 7:00 a. m. E. S. T., January 27, 1958.

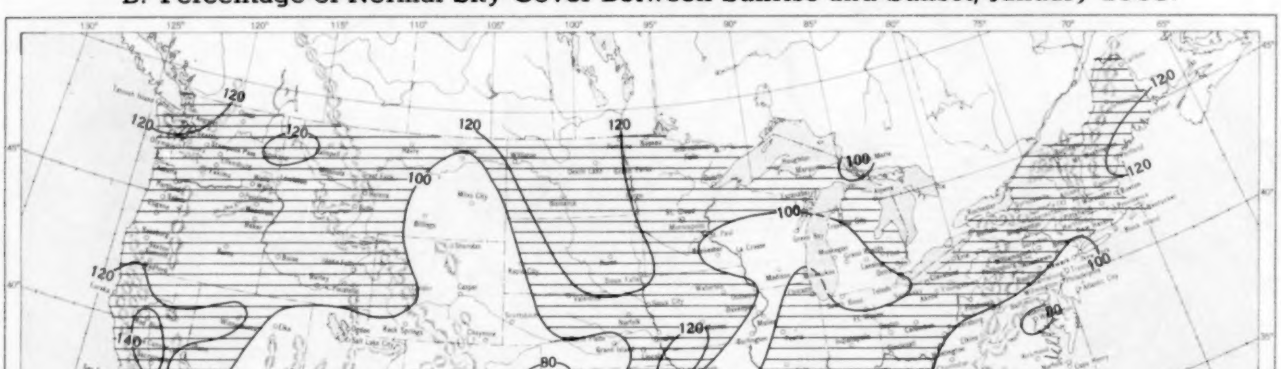


A. Amount of normal monthly snowfall is computed for Weather Bureau stations having at least 10 years of record.
B. Shows depth currently on ground at 7:00 a. m. E.S.T., of the Monday nearest the end of the month. It is based on reports from Weather Bureau and cooperative stations. Dashed line shows greatest southern extent of snowcover during month.

Chart VI. A. Percentage of Sky Cover Between Sunrise and Sunset, January 1958.

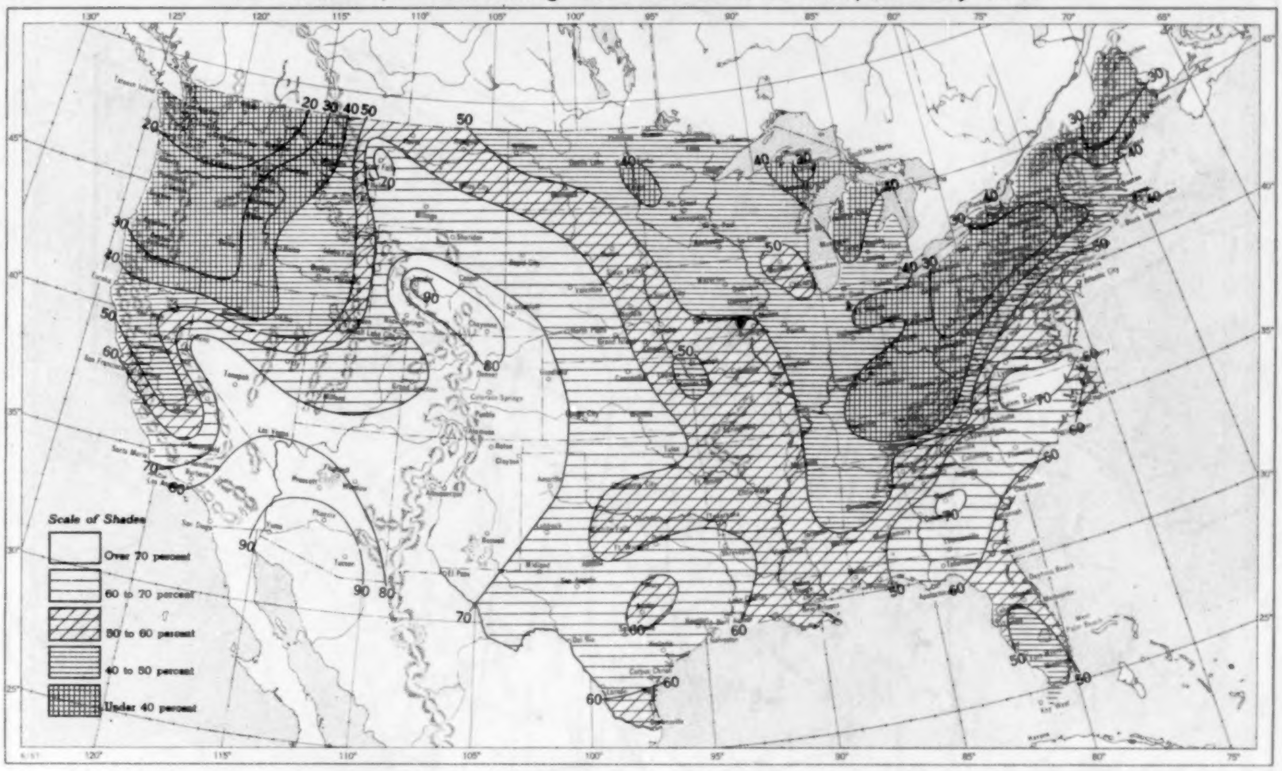


B. Percentage of Normal Sky Cover Between Sunrise and Sunset, January 1958.

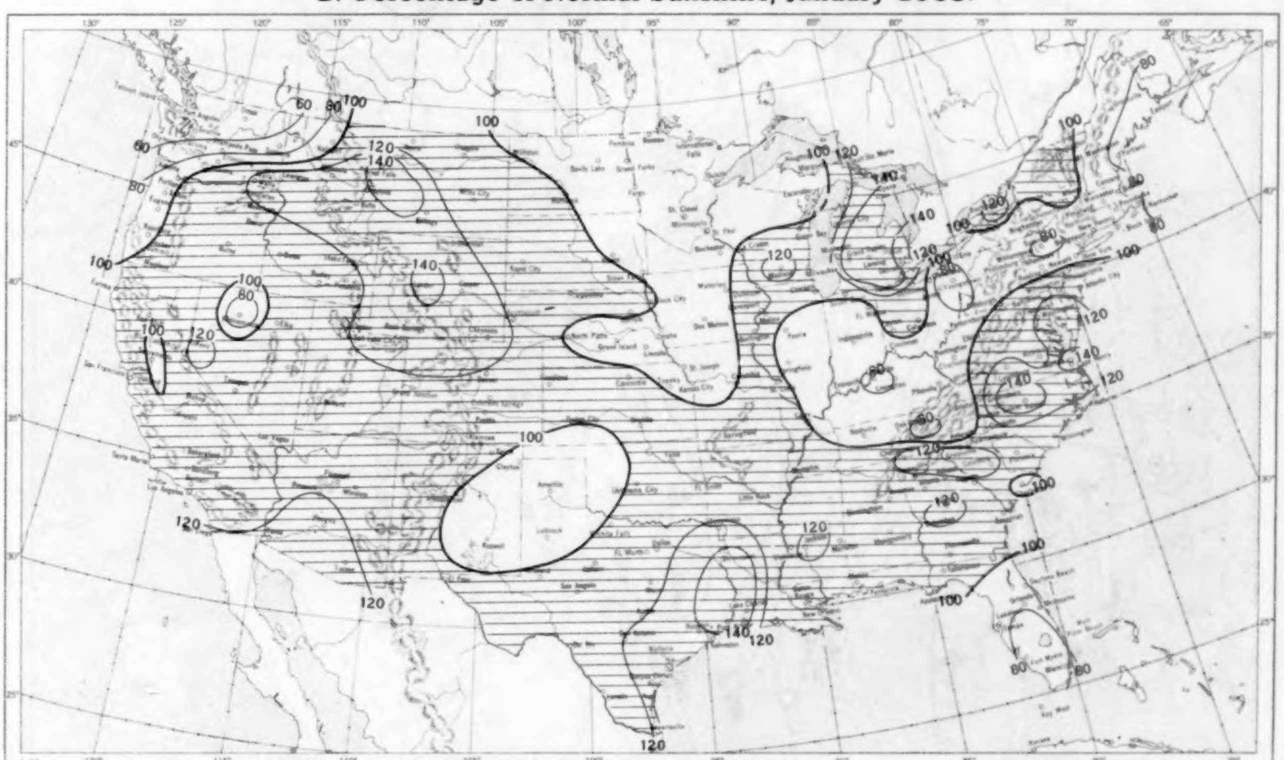


A. In addition to cloudiness, sky cover includes obscuration of the sky by fog, smoke, snow, etc. Chart based on visual observations made hourly at Weather Bureau stations and averaged over the month. B. Computations of normal amount of sky cover are made for stations having at least 10 years of record.

Chart VII. A. Percentage of Possible Sunshine, January 1958.



B. Percentage of Normal Sunshine, January 1958.



A. Computed from total number of hours of observed sunshine in relation to total number of possible hours of sunshine during month. B. Normals are computed for stations having at least 10 years of record.

Chart VIII. Average Daily Values of Solar Radiation, Direct + Diffuse, January 1958. Inset: Percentage of Mean Daily Solar Radiation, January 1958. (Mean based on period 1951-55.)

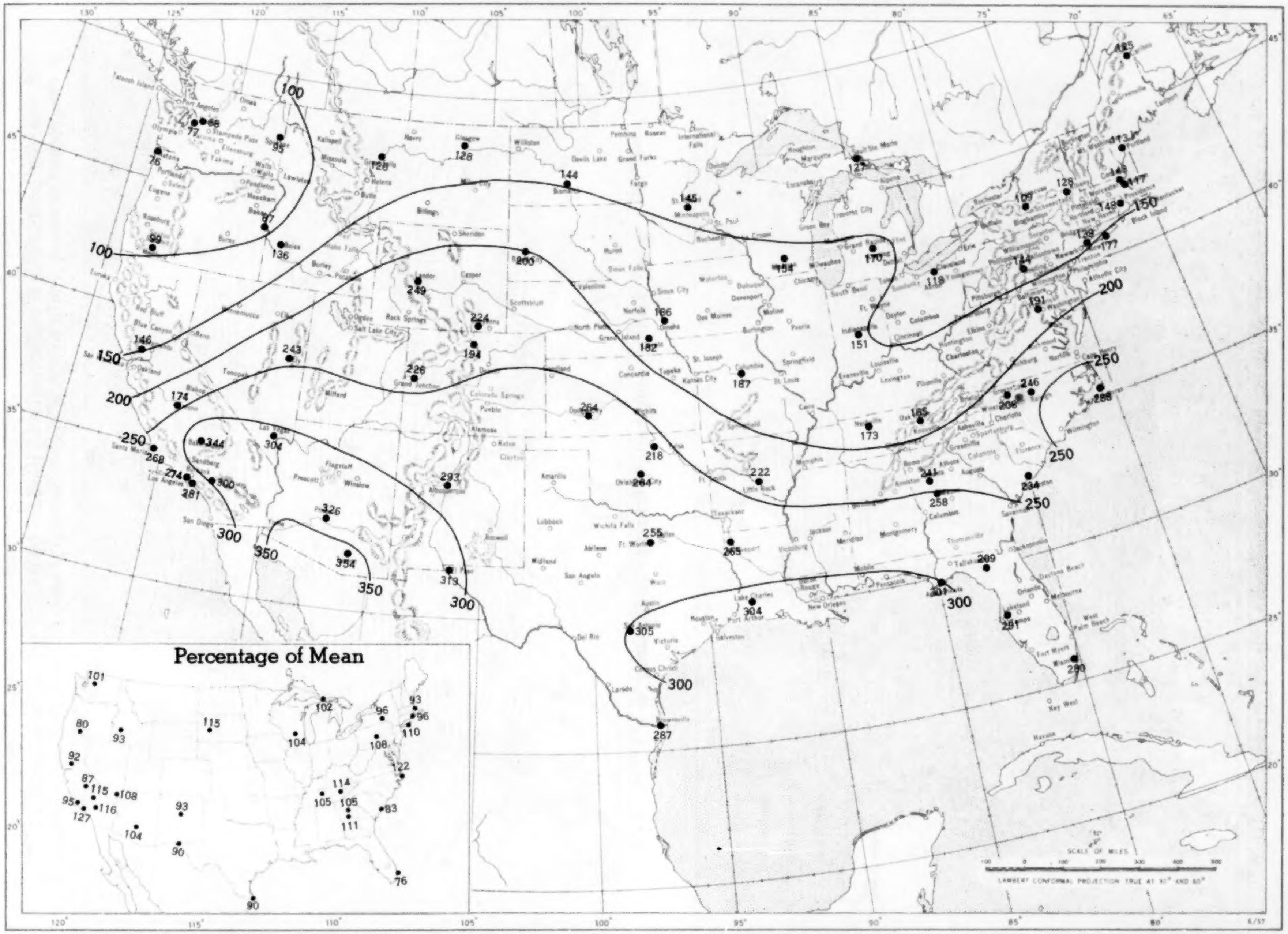
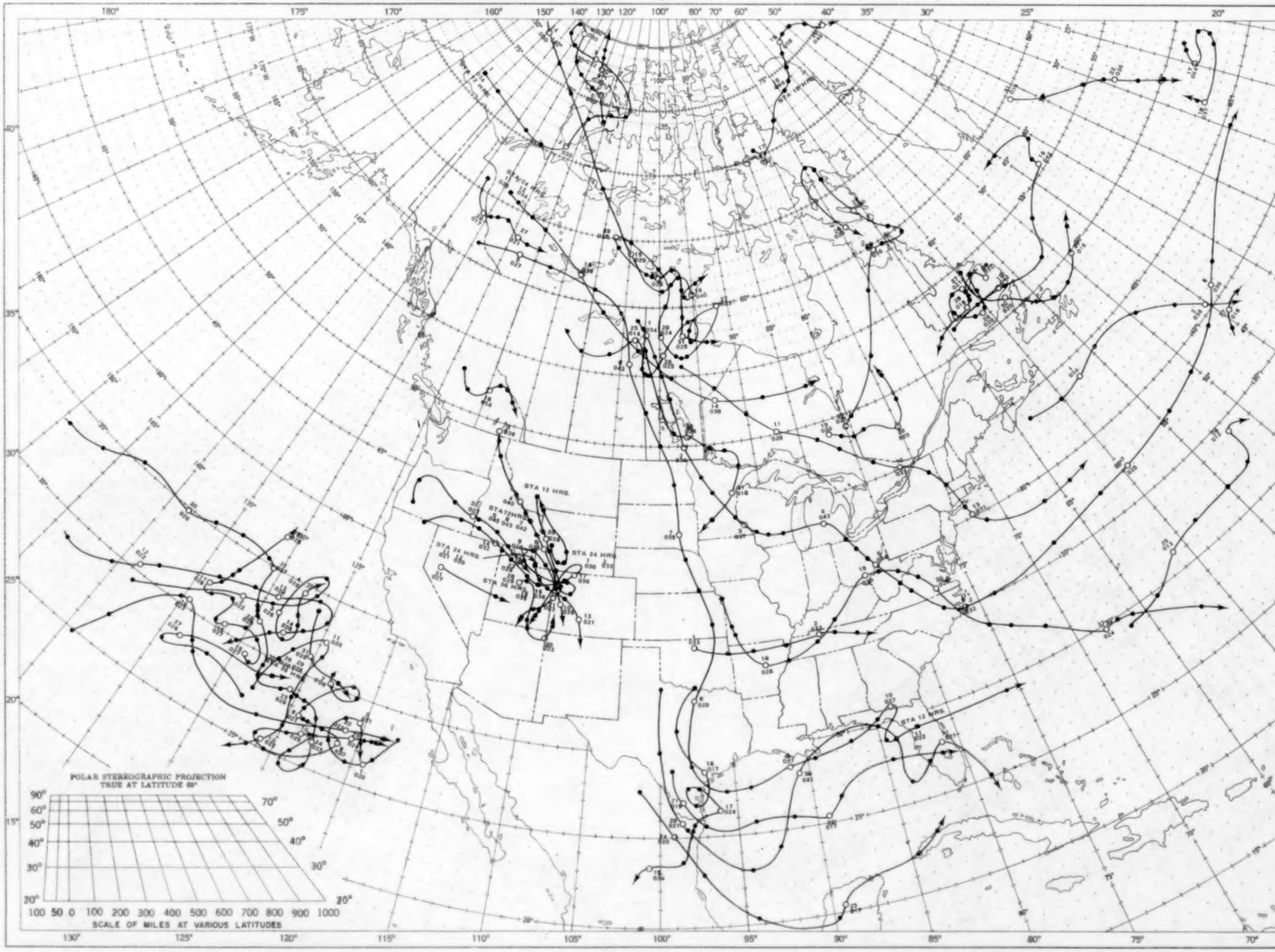


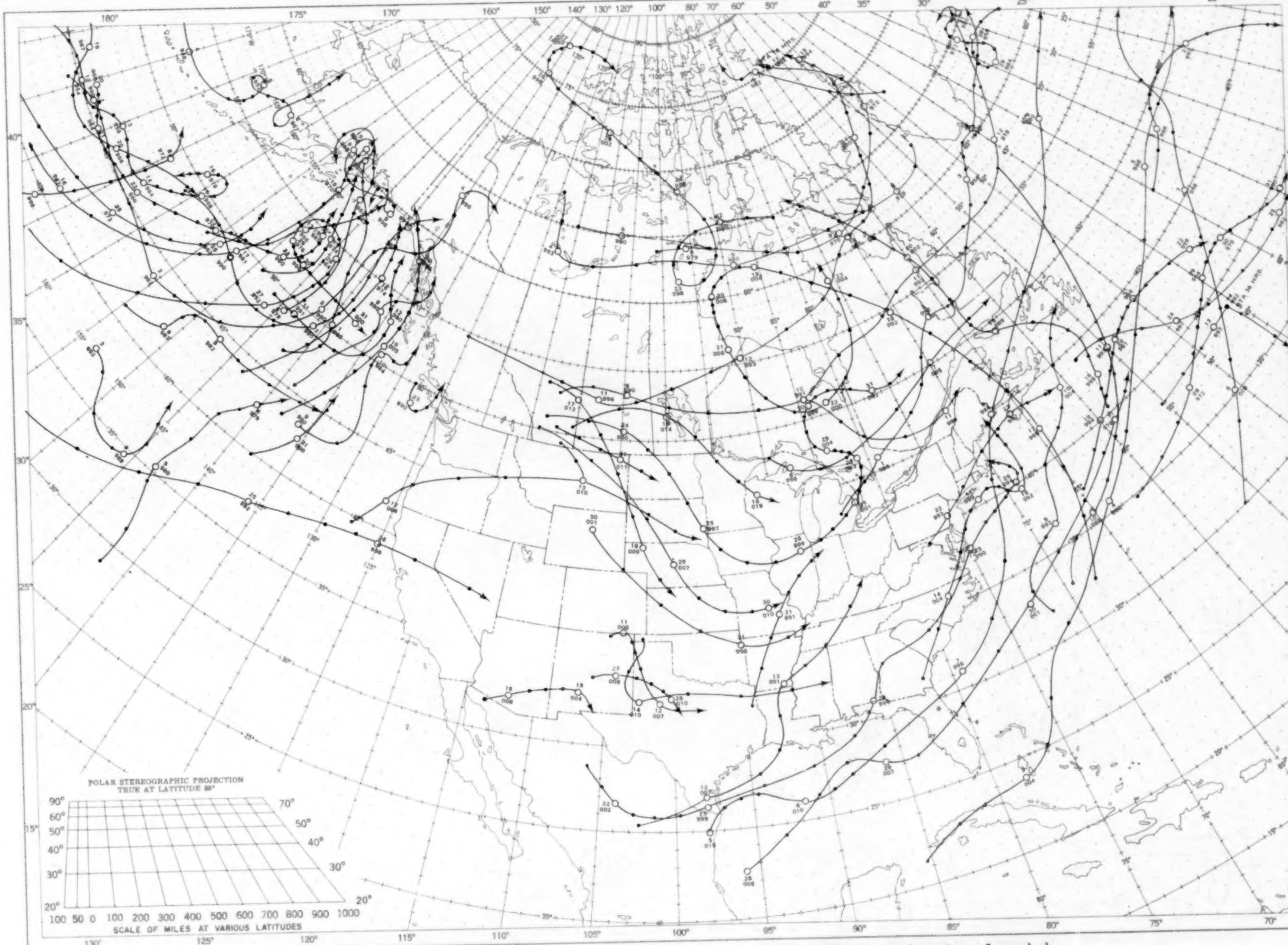
Chart shows mean daily solar radiation, direct + diffuse, received on a horizontal surface in langleys (1 langley = 1 gm. cal. cm. $^{-2}$). Basic data for isolines are shown on chart. Further estimates are obtained from supplementary data for which limits of accuracy are wider than for those data shown. The inset shows the percentage of the mean based on the period 1951-55.

Chart IX. Tracks of Centers of Anticyclones at Sea Level, January 1958.



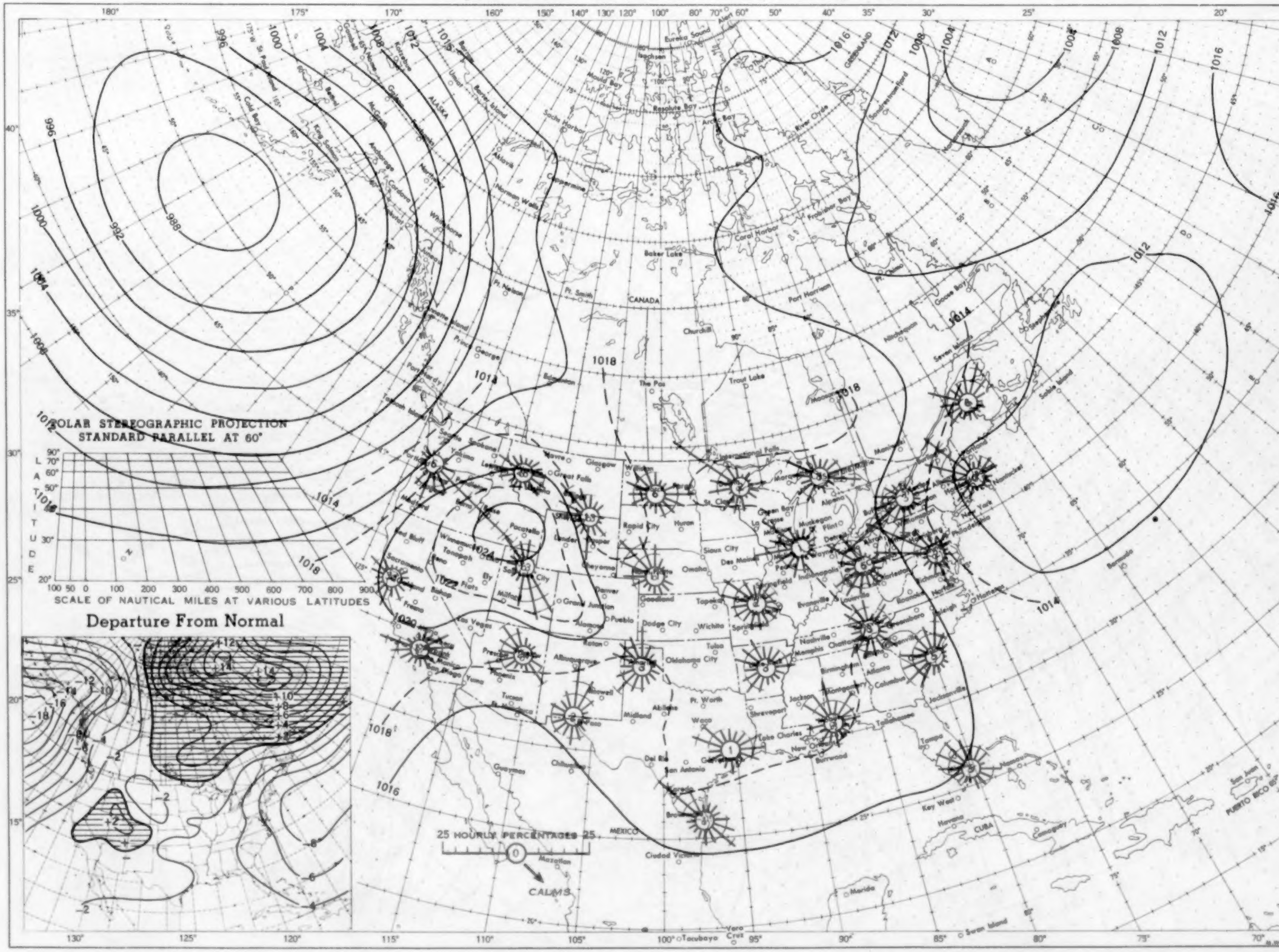
Circle indicates position of center at 7:00 a. m. E. S. T. Figure above circle indicates date, figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Squares indicate position of stationary center for period shown. Dashed line in track indicates reformation at new position. Only those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, January 1958.



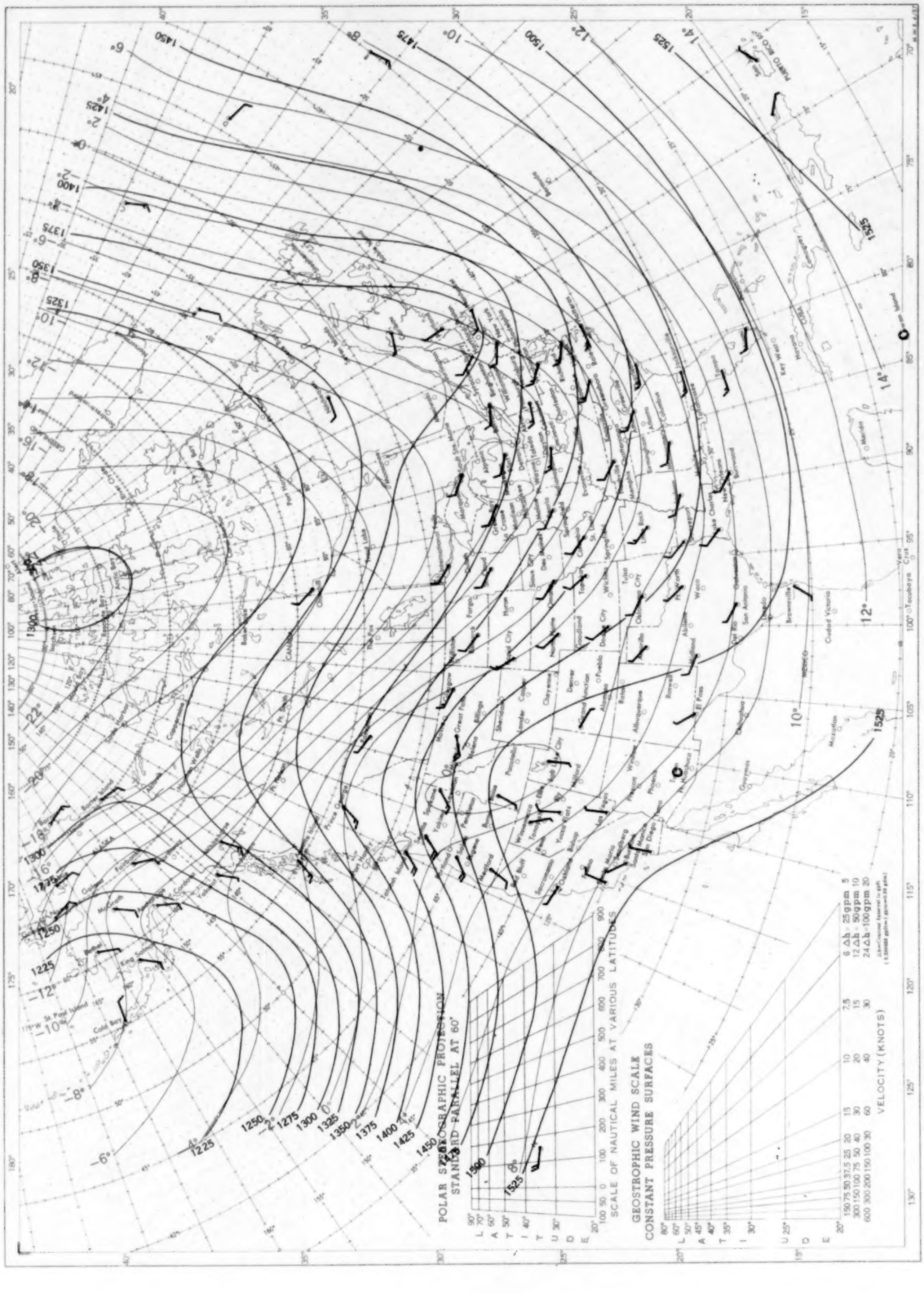
Circle indicates position of center at 7:00 a. m. E. S. T. See Chart IX for explanation of symbols.

Chart XI. Average Sea Level Pressure (mb.) and Surface Windroses, January 1958. Inset: Departure of Average Pressure (mb.) from Normal, January 1958.



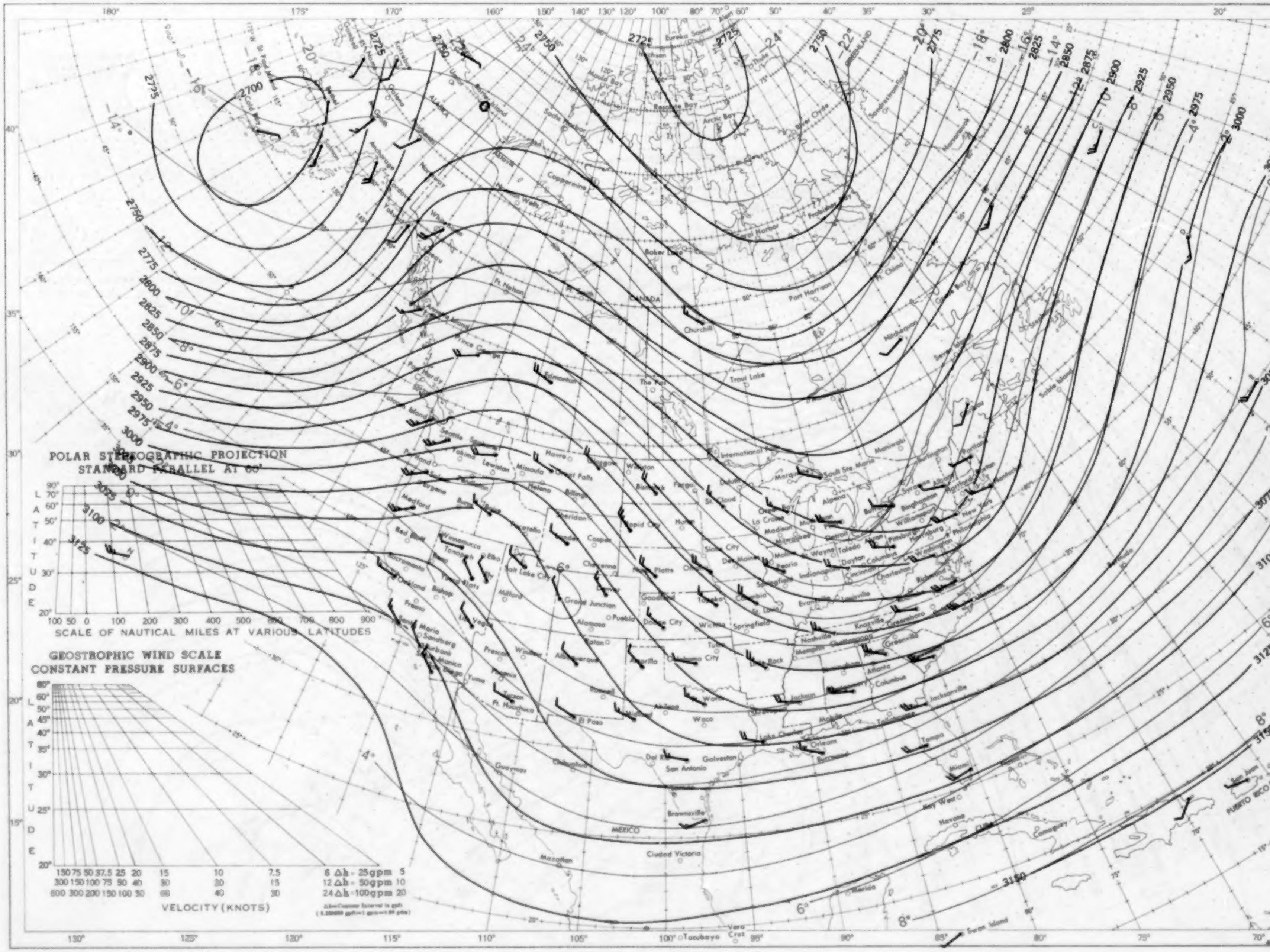
Average sea level pressures are obtained from the averages of the 7:00 a. m. and 7:00 p. m. E. S. T. readings. Windroses show percentage of time wind blew from 16 compass points or was calm during the month. Pressure normals are computed for stations having at least 10 years of record and for 10° intersections in a diamond grid based on readings from the Historical Weather Maps (1899-1939) for the 20 years of most complete data coverage prior to 1940.

Chart XII. 850-mb. Surface, 1200 GMT, January 1958. Average Height and Temperature, and Resultant Winds.



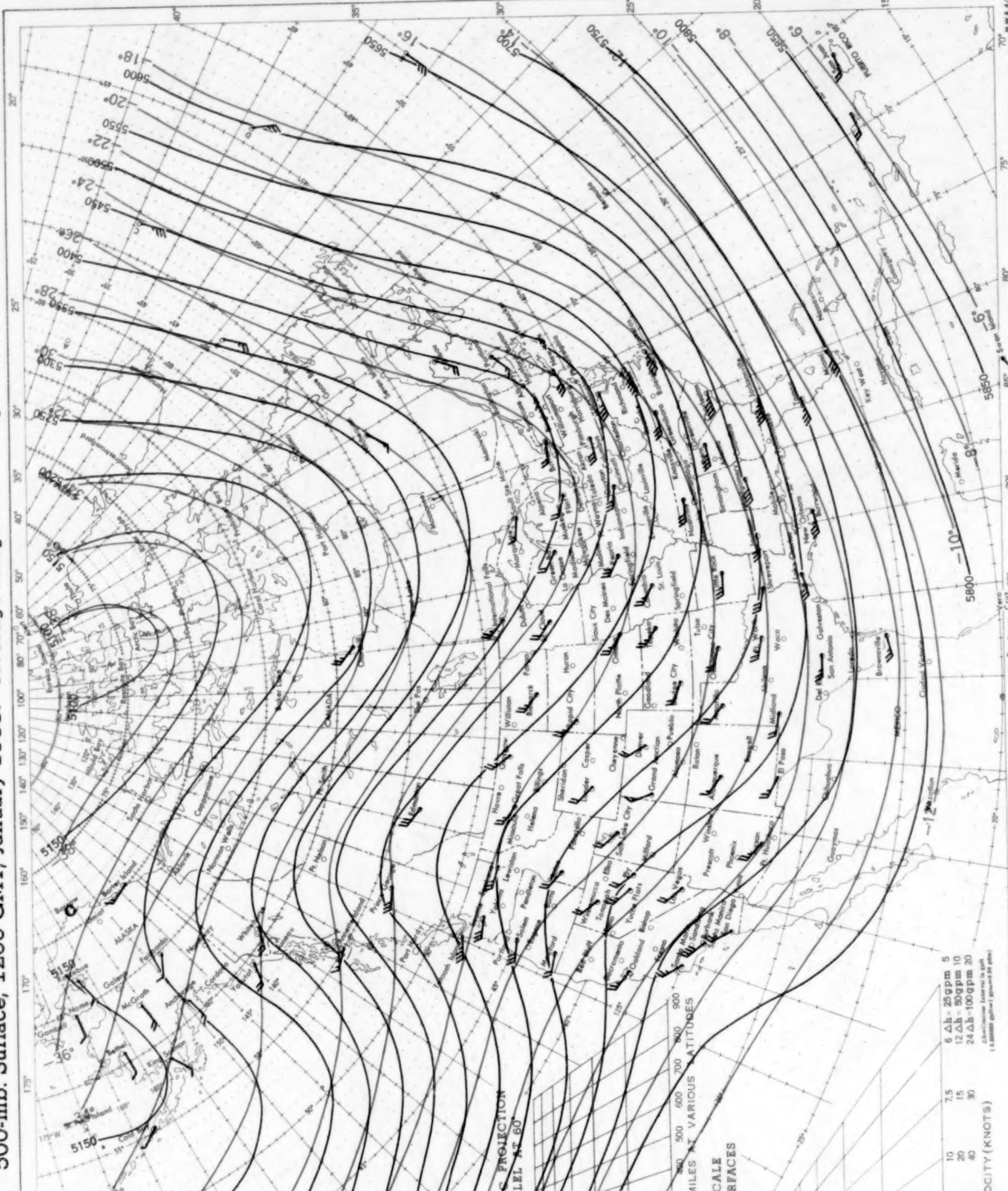
Height in geopotential meters (1 g.p.m. = 0.98 dynamic meters). Temperature in °C. Wind speed in knots; flag represents 50 knots, full feather 10 knots, and half feather 5 knots. All wind data are based on rawin observations.

Chart XIII. 700-mb. Surface, 1200 GMT, January 1958. Average Height and Temperature, and Resultant Winds.



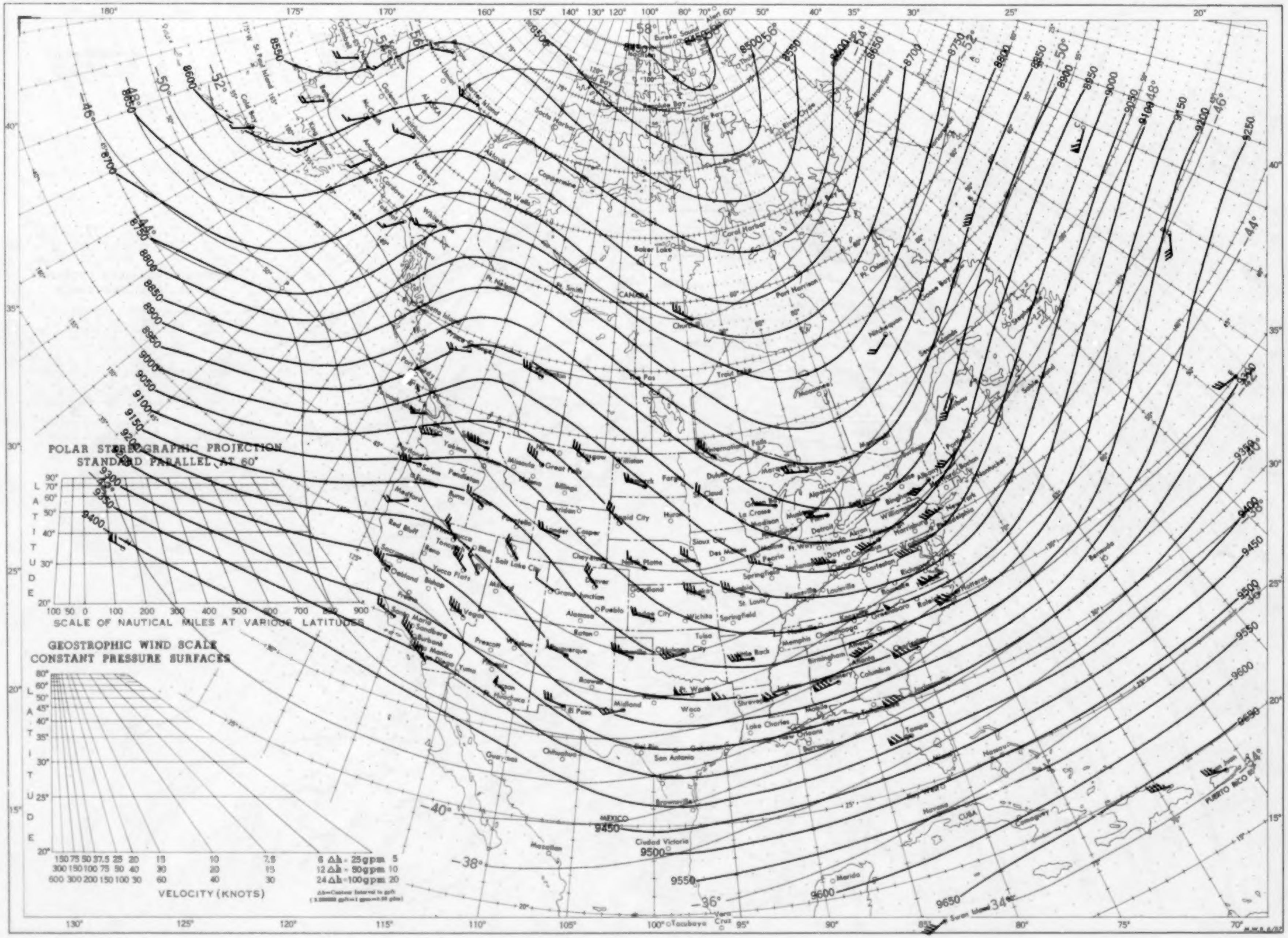
See Chart XII for explanation of map.

Chart XIV. 500-mb. Surface, 1200 GMT, January 1958. Average Height and Temperature, and Resultant Winds.



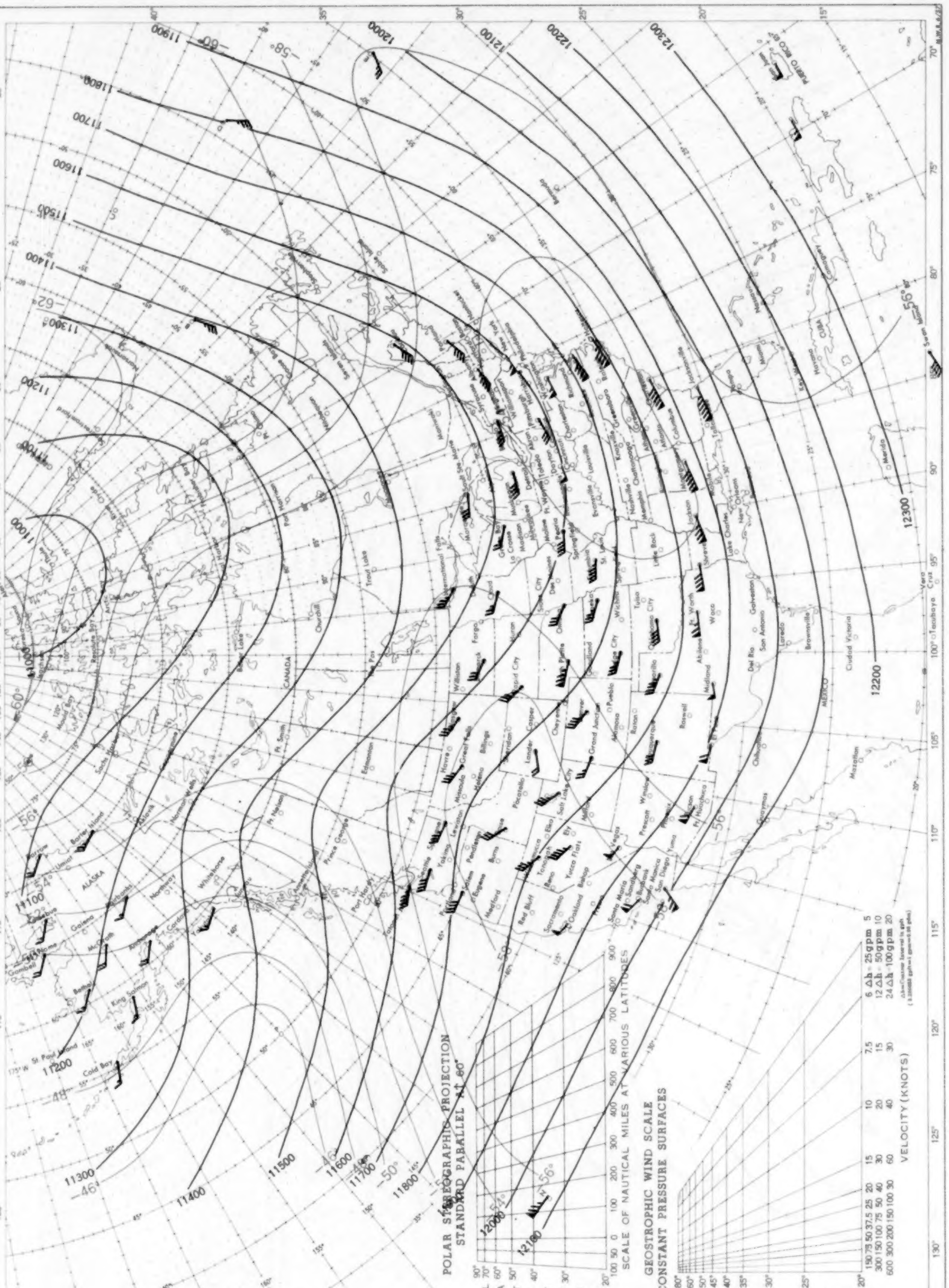
See Chart XII for explanation of map.

Chart XV. 300-mb. Surface, 1200 GMT, January 1958. Average Height and Temperature, and Resultant Winds.



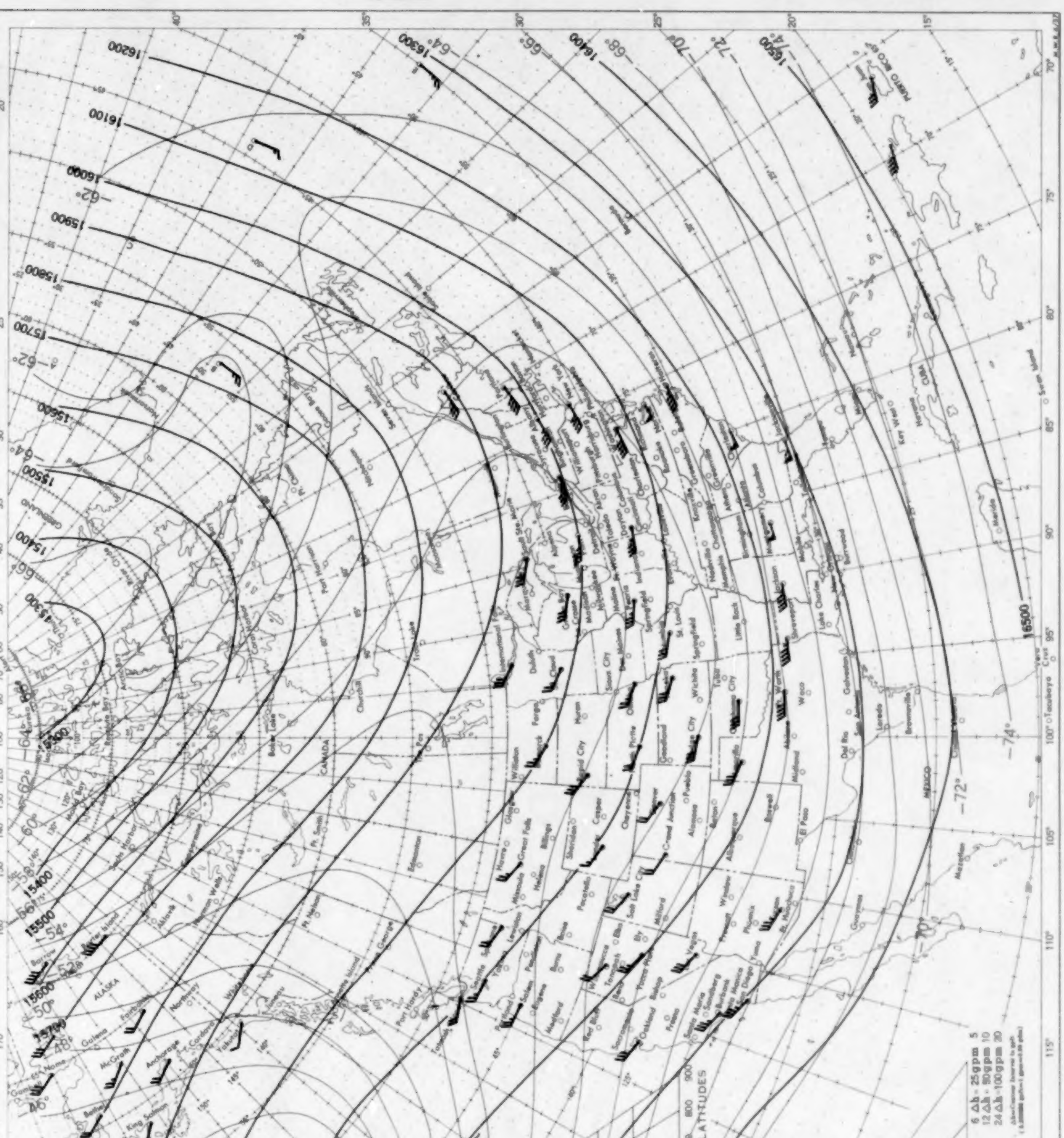
See Chart XII for explanation of map.

Chart XVI. 200-mb. Surface, 1200 GMT, January 1958. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map.

Chart XVII. 100-mb. Surface, 1200 GMT, January 1958. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map.